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MRP No.	Title	Author
752 AD 032643	Some Downstream Features Associated With Large Amplitude Troughs in Upper Air Circulation Patterns.	C.V. Smith A.G. Forsdyke
755 AD 032650	The Formation of New Anticyclones.	R.C. Sutcliffe
757	Meteorological Report on the Singapore Cloud Detection Trials.	R. Frost
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777 AD 032645	A Study of Depressions Crossing the Region of Labrador and the St. Lawrence Basin.	A.G. Forsdyke

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M.R.P. 802
15th October, 1953

R 5544

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MRIS

Control Number C.D.8063

M.R.P. 752
S.C.II/118
22nd September, 1952.

AD 32643

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee, London)."
M.R.P. No ..... by Mr.....
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METEOROLOGICAL RESEARCH COMMITTEE

Some downstream features associated with large amplitude troughs in upper air circulation patterns

by

C.V. Smith and A.G. Forsdyke (Meteorological Office).

Summary

The change in amplitude of troughs existing or forming downstream of a (primary) trough in the 500 mb. contour and 1000 - 500 mb. thickness patterns over America is examined for the four days subsequent to the first formation of the American trough.

Over the time interval considered, the association between a large amplitude primary and the presence, in a broad sector eastward of the primary, of a large amplitude downstream trough is found, in general, to be small. The association increases markedly however when a primary persists for two or three days; on a large proportion of such occasions the next downstream trough attains dimensions comparable with that of the primary. Persistence of this downstream trough would frequently appear to be directly related to persistence of the primary.

Certain incidental results concerning new trough formation are examined for evidence in support of particular aspects of barotropic flow theory. The modes of formation of downstream troughs are discussed in the Appendix.

1. Introduction

In the preparation of experimental extended (four-day) forecasts a qualitative assessment of the future modifications and developments of the 500 mb. flow pattern often decides the "model" or the broad scale evolution of surface features to be adopted. It has at times been argued in this connection that the formation of a large amplitude trough should lead to the development of a correspondingly large amplitude trough downstream. If such a process could be shown to take place within four days of the initial formation, say, of a trough over America, it would obviously have an important bearing on extended forecasts issued for the British Isles. The primary concern of the present work was the accumulation of some statistics which would demonstrate the general validity or otherwise of the above argument.

2. Procedure: Selection of Data

Circumpolar 500 mb. contour charts at 0300 G.M.T. each day for the year 1950 were examined for the formation of "simple" troughs over America. Such a trough was taken to be one in which the estimated central flow line in its southern portion approximated, as judged by eye, to one of a family of symmetrical curves,

/(parabolas),

(parabolas), and whose axis was tilted less than 20° to the meridian through the vertex of the central flow line. An example is shown in Fig.1. The selection of these initially simple primary troughs from the gamut available for discussion was made with the object of achieving a reasonable degree of uniformity in the cases studied. It was hoped that simple downstream patterns might result from simple primaries so that any "downstream effect" would be the more readily discernible; further, that any results obtained from the initial study of simple systems would apply in some helpful degree to more complex features.

A measure of the trough amplitude was obtained by noting the number of degrees of latitude from the vertex (L) of the central flow line to the intersection (I) given on the axis of the best fitting standard curve by the line joining the points of inflexion (P) of the central flow line on either side of the vertex. This measurement was made along the axis of the standard curve. The troughs were classified according to amplitude as follows:-

Amplitude A	15° latitude or more
B	10 - 14° latitude
C	5 - 9° latitude.

The longitudinal separation of any pair of troughs was measured from vertex to vertex.

In some cases, (e.g. Fig.1), the presence of minor disturbances in the general flow round the trough made it difficult or impossible to locate a point of inflexion on one side or the other of the trough. On these occasions, an estimate was made of the position which the point of inflexion would have occupied if the disturbances had been absent. This is obviously a subjective process. Since however the object was to categorise troughs into one of three broad classes, each covering a range of amplitudes, rather than to specify the trough amplitude precisely, such a subjective process should not affect the results materially.

The sector considered for downstream effects was that eastward of the primary trough to longitude 50°E.

Although simple troughs could be selected over America, those present or forming downstream were not always of the same simple structure. As far as possible similar measurements were made on these features as on the primary troughs. The restriction regarding tilt of axis was dropped.

The behaviour of troughs present downstream on the day of first formation of the primary American trough, and also the primary itself, was studied for the four days following the first formation of the primary. A study was made also of new troughs i.e., those which were not present downstream in the sector on the day of formation of the primary.

The results of these investigations are tabulated and discussed in paragraphs 3 - 5 below. The results of a similar investigation for troughs in the 1000 - 500 mb. thickness pattern are given in paragraph 6.

In the tables, distinction is made between primary troughs of amplitude (A) and (B), and the cases shown are independent. Many more cases of primary (B) occurred than shown but in order to avoid the inclusion of some data in both parts of the tables, only those primary (B) troughs were considered which formed at least four days before the formation, or at least four days after the degeneration of a primary trough of amplitude (A).

3. Amplitude of downstream contour troughs

(a) The data

Within the sector eastward of the primary to 50°E. up to three downstream troughs were present on some occasions. Table I is a frequency table

giving the amplitude of these downstream troughs (taken in order from W. to E.), on the day of formation (D0) of the primary and of the amplitude of these same troughs on the four subsequent days (D1 to D4). The cases of large (A) and moderate (B) amplitude primary troughs are tabulated separately.

X is the number of the downstream troughs considered on D0 which subsequently degenerated into complex and indefinite features and which were consequently no longer classifiable, or whose amplitude fell to less than 5° latitude (the minimum for classification) or which passed eastward out of the sector beyond 50°E.

TABLE I

Amplitude of downstream troughs

Primary trough amplitude (A) (27 cases)

First downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	1	2	4	3	2
B	17	14	7	10	12
C	9	7	9	7	7
X	-	4	7	7	6
Total	27				

Second downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	1	1	1	1	2
B	7	4	2	2	3
C	14	13	9	4	1
X	-	4	10	15	16
Total	22				

Third downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	2	1	1	2	1
B	1	3	1	-	-
C	2	-	1	-	-
X	-	1	2	3	4
Total	5				

Total of all downstream troughs.

Amplitude	D0	D1	D2	D3	D4
A	4	4	6	6	5
B	25	21	10	12	15
C	25	20	19	11	8
X	-	9	19	25	26
Total	54				

Primary trough amplitude (B) (21 cases)

First downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	4	3	2	2	3
B	10	8	7	7	4
C	6	6	6	6	5
X	-	3	5	5	8
Total	20				

Second downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	-	1	1	1	-
B	7	6	6	2	4
C	6	4	2	3	1
X	-	2	4	7	8
Total	13				

Third downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	1	-	-	-	-
B	1	2	1	1	1
C	-	-	-	-	1
X	-	-	1	1	-
Total	2				

Total of all downstream troughs.

Amplitude	D0	D1	D2	D3	D4
A	5	4	3	3	3
B	18	16	14	10	9
C	12	10	8	9	7
X	-	5	10	13	16
Total	35				

There was one occasion with primary (B) on which there was no downstream trough on D0 falling within our classification.

(b) Conclusions from Table I

The following apply only in relation to those downstream features which were already present on D0; they may be modified by considerations relating to the formation of new troughs discussed in paragraph 5.

(i) There is little tendency for a large or medium amplitude primary trough to be associated either on D0 or during the next four days with a large amplitude (A) downstream trough; on any one day such a feature is found with only about one in five of the primary troughs.

/(ii)

(ii) On D0 the total number of downstream troughs of amplitude (B) relative to the total number of primaries (A) or (B) is 80% or more. (Since, however, more than one of these downstream troughs may be present on the same day, the percentage of occasions with a downstream trough of amplitude (B) may be somewhat less.) This proportion falls off considerably during the following four days. The partial recovery on D3 and D4 with primary (A) may perhaps be connected with the persistence of the primary as discussed in paragraph 8.

4. Change of amplitude of downstream contour troughs

(a) The Data

The individual changes of amplitude of troughs present downstream of the primary on the day of its formation, D0, are shown below in Table II, for each of the four days D1 to D4.

The trough amplitude on D0 is taken as the zero level from which subsequent increase or decrease is reckoned when the trough amplitude passes from one range to another on the scale A,B,C. Changes of amplitude range from A to B, B to A, for example, are denoted by minus one unit and plus one unit respectively. X has the meaning indicated in Table I. The reasons for which troughs were placed in category X are analysed in Table IIa.

TABLE II

Change of amplitude of downstream troughs

Primary trough amplitude (A)						Primary trough amplitude (B)					
Units of Change	D0	D1	D2	D3	D4	Units of Change	D0	D1	D2	D3	D4
+2	-	-	1	1	2	+2	-	1	-	1	-
+1	-	3	6	9	9	+1	-	2	4	2	4
0	54	37	19	13	10	0	35	20	15	11	7
-1	-	5	8	6	7	-1	-	7	4	6	7
-2	-	-	1	-	-	-2	-	-	2	2	1
X		9	19	25	26	X	-	5	10	13	16

TABLE II(a)

Analysis of reasons for which troughs were no longer classified and so placed in category X defined above.

	Primary trough amplitude (A)				Primary trough amplitude (B)			
	D1	D2	D3	D4	D1	D2	D3	D4
Trough amplitude less than 5°	5	8	11	11	4	7	9	12
Trough becomes complex	1	6	6	7	1	2	2	2
Trough passes from sector	3	5	8	8	-	1	2	2
Total	9	19	25	26	5	10	13	16

(b) Conclusions and remarks

(i) The proportion of downstream troughs which preserve their amplitude and simple structure shows a steady falling off over the period, to about one in five on the fourth day.

(ii) The proportion of troughs in category X shows a steady increase over the period, from about one in six on D1 to one in two on D4.

(iii) The proportion of troughs which undergo an increase of amplitude is small, amounting at most on any one day to one in five.

/(iv)

(iv) In preparing Table II(a) it was often difficult to distinguish between troughs of amplitude less than 5° and those which became complex. Little significance should therefore be attached to the relative numbers in these categories. The high proportion of troughs passing out of the sector with primary (A) as compared with primary (B) might be thought to be significant, indicating either greater mobility or a greater initial longitudinal separation. The latter is suggested by a calculation of the mean longitudinal separation of the primary and first downstream troughs, which gives about 75° for primary (A) and 65° for primary (B).

5. New downstream contour troughs

(a) The Data

The amplitude and frequency of new troughs (not present on D0), present in the sector eastward of the primary to 50°E. on the four days, D1 - D4, following the formation of a primary on D0 is shown in Table III. N is here used to denote the number of occasions on which no new features occurred. Any apparent discrepancy between the total of new features and the total of occasions on which new features formed is explained by the appearance of more than one new feature on certain days.

TABLE III

Amplitude of new downstream troughs

Primary trough amplitude (A)(27 occasions)

Amplitude	D1	D2	D3	D4
A	-	1	1	3
B	1	1	2	3
C	10	7	5	7
N	16	21	20	15

Primary trough amplitude (B)(21 occasions)

Amplitude	D1	D2	D3	D4
A	1	1	1	1
B	-	-	2	2
C	3	5	3	4
N	17	15	15	14

(b) Conclusions and remarks

(i) There is little tendency for the formation of a large (A) or moderate (B) amplitude primary trough to be followed by the appearance of a new downstream trough of large or moderate amplitude.

(ii) With primary trough (A) or (B), no new features are present at all on 55% or more of occasions.

(iii) The number of new features of amplitude (A) or (B) is small, but as might be expected, shows a slight cumulative effect over the period. The falling off between D1 and D4 of new features (C) associated with primary (A) appears peculiar at first sight suggesting perhaps that all the new features were formed on D1 and no more thereafter; a reference to the original data however shows that most of the new small amplitude troughs were short-lived and that some such new features formed on each of the four days.

(iv) Combining our results for new features with those for features already in existence on D0 (Tables III and I) we notice that the total number of large amplitude (A) downstream troughs tends to increase over the period in the case of primary (A). However, the proportion in relation to the number of primaries (A) or (B) does not reach more than about 30% on any day.

(v) The formation of new features of amplitude (B) or (C), (Table III), is insufficient to balance the loss of existing features, (Table I). Table IIa shows that the net loss is due mainly to troughs becoming complex or of small amplitude.

6. 1000 - 500 mb. thickness troughs

In our step by step production of the sequence of prebaratic charts for an extended forecast, the 1000 - 500 mb. thickness isopleths rather than the 500 mb.

contours have been the principal aid in the treatment of surface features. It is of interest therefore to know whether the conclusions drawn from the study of 500 mb. contours may be applied also to the 1000 - 500 mb. thickness isopleths.

Taking the 1000 - 500 mb. thickness charts at 0300 G.M.T. each day for the same year, 1950, the procedure applied to the 500 mb. contour charts was repeated. The results are given in Table IV - VII. Some conclusions are given in paragraph 7.

Table IV gives the data for troughs in the thickness pattern corresponding to Table I for troughs in the contour pattern. Similarly Table V corresponds to Table II and Table VI corresponds to Table III.

A comparison of the number of "simple" primary troughs obtained from the two sets of isopleths studied is interesting.

Primary Trough (A) Contours 27, Thickness 27.

Primary Trough (B) Contours 21, Thickness 26.

The two sets of data were obtained independently but later comparison showed that simple troughs in the contours could be associated with simple troughs of the same amplitude in the thickness pattern on about 60 % of occasions. Simple troughs of amplitude (A) or (B) in the contours could be associated with simple troughs of amplitude (A) or (B) in the thickness pattern on about 85% of occasions.

7.

TABLE IV

Amplitude of downstream (thickness) troughs

Primary trough amplitude (A)(27 cases)

First downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	4	4	3	4	6
B	7	7	7	7	5
C	16	10	9	7	4
X	-	6	8	9	12
Total	27				

Second downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	-	2	3	4	3
B	15	13	10	8	5
C	7	2	3	2	2
X	-	5	6	8	12
Total	22				

Third downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	2	1	2	1	1
B	1	2	-	-	-
C	1	1	1	1	-
X	-	-	1	2	3
Total	4				

Total of all downstream troughs.

Amplitude	D0	D1	D2	D3	D4
A	6	7	8	9	10
B	23	22	17	15	10
C	24	13	13	10	6
X	-	11	15	19	27
Total	53				

Primary trough amplitude (B)(26 cases)

First downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	1	1	-	2	2
B	8	11	13	6	7
C	16	11	10	10	5
X	-	2	2	7	11
Total	25				

Second downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	-	2	2	1	1
B	7	3	3	5	5
C	11	9	5	6	6
X	-	4	8	6	6
Total	18				

Third downstream trough.

Amplitude	D0	D1	D2	D3	D4
A	1	1	1	1	1
B	5	1	2	-	-
C	3	5	2	2	1
X	-	2	4	6	7
Total	9				

Total of all downstream troughs.

Amplitude	D0	D1	D2	D3	D4
A	2	4	3	4	4
B	20	15	18	11	12
C	30	25	17	18	12
X	-	8	14	19	24
Total	52				

There was one occasion with primary (B) on which there was no downstream trough on D0 falling within our classification.

TABLE V

Change of amplitude of downstream (thickness) troughs

(See Table II for explanation).

Primary trough amplitude (A)

Units of Change	D0	D1	D2	D3	D4
+2	-	-	1	3	2
+1	-	10	9	8	10
0	53	26	20	18	10
-1	-	6	8	5	4
-2	-	-	-	-	-
X	-	11	15	19	27

Primary trough amplitude (B)

Units of Change	D0	D1	D2	D3	D4
+2	-	-	-	3	2
+1	-	11	12	4	6
0	52	24	21	21	16
-1	-	9	5	4	4
-2	-	-	-	1	-
X	-	8	14	19	24

TABLE V (a)

Analysis of reasons for which troughs were no longer classified and so placed in category X defined above

	Primary trough amplitude(A)				Primary trough amplitude(B)			
	D1	D2	D3	D4	D1	D2	D3	D4
Trough amplitude less than 5°	6	10	11	14	5	7	8	13
Trough becomes complex	3	3	5	8	2	4	4	4
Trough passes from sector	2	2	3	5	1	3	7	7
Total	11	15	19	27	8	14	19	24

TABLE VI

New downstream (thickness) troughs

Primary trough amplitude(A)(27 cases)

Amplitude	D1	D2	D3	D4
A	-	-	-	1
B	1	2	2	3
C	1	2	7	5
N	25	23	19	19

Primary trough amplitude(B)(26 cases)

Amplitude	D1	D2	D3	D4
A	-	-	1	-
B	-	4	2	4
C	9	11	10	6
N	18	13	15	17

Remarks

Although there are differences between the corresponding tables for contour and thickness troughs these are mainly differences of detail and their significance is probably small. The most notable difference is the greater number in the case of thickness as compared with contour troughs, of large amplitude downstream features with primary amplitude (A). (Tables I and IV). Broadly, the conclusions which we have arrived at for contour troughs apply equally to thickness troughs, and call for no further comment.

8. Persistent primary troughs in the contour patterns

(a) The Data

The investigation so far has been concerned with the events following the first formation of a simple large or moderate amplitude trough without taking into account whether or not this primary was more than a passing feature or subsequently retained its simple structure.

The number of cases of persistent simple troughs i.e., those which retained their amplitude and simple structure as defined in paragraph 2 for more than

one day, in the year 1950, was found to be insufficient for any definite conclusions to be drawn; the initially simple troughs were frequently present with a more complex structure or with smaller amplitude though still as important features, throughout the entire four day period.

From this it was concluded that the initial criteria were too restrictive for anything practical to result from their retention when persistence was to be considered. The following procedure was consequently adopted.

The behaviour of the 27 troughs having amplitude (A) and a simple structure on D0 was followed over the four day period D0 - D4. Since troughs of amplitude (A) tend to be the extreme cases of oscillatory flow the primary was now said to persist when the primary trough amplitude fell to (B). Again if some complexity was present for one day only, no departure from our criteria for simple troughs being observed on the preceding or following day, the primary was still said to persist. (A frequent source of complexity was the development of a perturbation on the western side of the trough which increased in amplitude and was effective in causing the trough to retrogress). Downstream, only the first trough was considered. This was taken as the next downstream trough with which the main flow of the primary was linked and sometimes differs from the first downstream trough with longitude as the sole criterion. When the primary underwent some complication so that it could no longer be classified, the behaviour of the downstream trough initially studied was followed until such time as a new first downstream trough was established, provided also that the primary was regenerated; if the identity of the primary was lost altogether, the downstream trough initially studied was followed until the end of the period.

Table VII shows the occasions, day by day, on which the first downstream trough had amplitude (A) and amplitude (A) or (B) in relation to the persistence of the primary. The cases shown are independent, e.g. primary troughs persisting for four days have not been included in the cases of primaries persisting for less than this time.

TABLE VII

Downstream troughs in relation to persistence of the primary

Primary trough present	Number of cases of primary	First downstream trough									
		Amplitude (A)					Amplitude (A) or (B)				
		D0	D1	D2	D3	D4	D0	D1	D2	D3	D4
D0 or D0 - D1	10	1	-	1	-	1	7	7	4	3	3
D0 - D2 or D0 - D3	8	1	-	2	3	3	4	5	4	7	7
D0 - D4	9	2	2	2	4	3	8	5	8	9	8

(b) Conclusions

(i) Taking a mean over the five days D0 - D4, there is a steady increase in the proportion of occasions on which the first downstream trough has amplitude (A), with persistence of the primary. This proportion is negligible with primary persistence of the order one day but increases to about one third of occasions with primary persistence of four days. When the downstream trough is of amplitude (A) or (B) the corresponding figures increase from about one half to five sixths of occasions.

(ii) With the longer persisting primaries there is a tendency for the relative number of downstream features of amplitude (A) or (B) to increase with time over the period D1 to D4.

(c) Consideration of additional cases of persistent primary troughs

Broadly it would seem from Table VII that persistence of the downstream trough might be related to persistence of the primary. In an attempt to elucidate this point a return was made to the original more stringent conditions of primary trough classification and this involved an examination of all available data (from September 1948 to March 1952) to enable sufficient cases for discussion to be obtained.

Conditions for persistence were taken now as follows:-

The primary should have amplitude (A) and simple structure on D0.

The introduction of some complexity or a decrease in primary amplitude to (B) for one day only was permissible provided that on the preceding and succeeding day the primary had amplitude (A) and a simple structure.

Downstream, as previously, only the behaviour of the first downstream trough with which the primary flow was linked, was considered.

TABLE VIII

Downstream troughs in relation to persistence of the primary (extended data)

Primary trough present	No. of cases of primary	No. of occasions when first downstream trough had amplitude (A)				
		D0	D1	D2	D3	D4
D0 - D1	15	3	3	5	2	2
D0 - D2	8	1	1	1	4	3
D0 - D3 or D0 - D4	7	1	4	3	5	7

Consideration of Table VIII leads to the following conclusions which in so far as they apply are in good agreement with paragraph 8(b).

(i) With increasing persistence of the primary there is an increasing tendency for the occurrence of large amplitude downstream features.

(ii) With primaries which are present for one or two days, the occurrence of downstream troughs of large amplitude reaches a maximum about D2 or D3 and then falls off again suggesting that the dissipation of the primary trough tends to be followed within about two days by dissipation of the downstream trough. Extension of the period to D5 and D6 indicates that the behaviour of downstream troughs associated with the more persistent primaries is in keeping with this suggestion.

9. Remarks on theory.

Although interaction between synoptic systems in different longitudes must have been recognised for a long time the conception, specifically, of downstream effects derives directly from recent work on long-wave flow patterns especially against the background of barotropic or equivalent barotropic theory. Barotropy implies to a close approximation conservation of absolute vorticity, and "constant vorticity trajectories" were introduced into forecasting (in America) as indicating where new troughs or ridges might form. It was then noticed that new oscillations sometimes emerged so distant from the "primary" as not to be explained by any advection of vorticity: a wave-like process of propagation was inferred. But theory had shown that the long-wave phase velocity was less than the mean zonal current and if the influence was to travel faster than by advection it must be by some wave velocity greater than the current velocity. Rossby (1945) suggested that the process was one of energy dispersion and showed, with various assumptions, that the group velocity should in fact be greater than the current velocity. Apart from affording a plausible explanation of developments not otherwise readily interpreted "barotropically", it is not clear what value is claimed for the idea of "dispersion" in synoptic meteorology and the present statistics do not afford much evidence.

In the first place it may be that the question is begged by speaking of a "primary" trough and implying that subsequent formations downstream arise consequentially. There is rarely evidence to prove that a downstream oscillation would not have arisen just as well without the presence of the primary and if, as the work of Eady (1950) and others implies, long-wave oscillations may arise as a result of "instability" there is no reason to look to "dispersion" for an

explanation: indeed, "dispersion" seems to be a useful concept only in so far as barotropic theory is invoked.

In the movement of a wave packet with a group velocity one would expect to see simultaneously a number of oscillations with the maximum amplitude being transferred forward from one trough-ridge to the next downstream. Our data do not suggest such a process but rather the reverse. The development of a large amplitude downstream trough is highly probable only when the "primary" trough is itself maintained; the impression is not one of energy transfer with a "group velocity" but of a local energy source such as only baroclinic theory can provide for.

If we assume that the emergence of a new trough is a downstream effect the speed of propagation may be estimated by dividing the longitudinal separation of "primary" and "consequent" by the time interval between their appearance. This has been done, as follows:-

For each new feature we evaluate the parameter

$$P = \frac{L_O - L_N}{D_N - D_O}$$

where L_O = longitude of primary on Day 0 (D0)

L_N = longitude of new contour trough on the day of its first appearance (DN).

The results are shown in Table IX which gives the frequency of occurrence of certain ranges of values of P.

TABLE IX

Frequency table of the ratio between longitudinal separation and time of formation of primary and new downstream contour troughs in degrees of longitude per day

P. Degrees per day	0-9	10-14	15-19	20-24	25-29	30-34	35-39	40-44	45-49	⩾50	Total
No. of cases (Primary A)	0	2	2	1	3	1	2	1	0	10	22
No. of cases (Primary B)	0	0	1	3	0	1	0	0	2	4	11
Total	0	2	3	4	3	2	2	1	2	14	33

The scatter is very large indeed. Since the mean westerly zonal current at 500 mb. over the period and half-hemisphere varied between 5° and 20° per day it is clear that the majority of the new troughs were not the result of simple advection. Rossby (1949) suggests that the maximum dispersive increment is about 18° per day but, allowing for this and extending the propagation speed to 40 degrees per day, only half the occasions are accommodated. From this point of view about half the occurrences might be dispersion effects (but of course are not necessarily such) and about half almost certainly are not.

Our data do not therefore lend much support to the dispersion hypothesis and certainly do not encourage the practical use in forecasting of formulae based upon it. This does not mean however that downstream effects do not occur, certainly on occasions the development of one oscillation is clearly linked with an earlier disturbance elsewhere, but the mechanism of the linkage probably involves the processes of baroclinic development.

/10.

10. Conclusions

(i) The suggestion which led to the investigation of this report namely that a large amplitude trough over America tends to be associated with or followed by a similar feature downstream has found only limited support; taking into account existing and newly forming features, it appears that such a large amplitude downstream trough occurs with not more than about one in three of the primary troughs.

(ii) If the primary trough is persistent (while retaining its "simple" structure) the occurrence of large amplitude downstream troughs becomes relatively much more frequent. This increase takes place towards the end of the four day period following the formation of the primary. However the occurrence of these persistent large amplitude primaries is rare (7 cases in $3\frac{1}{2}$ years). An initially large amplitude primary more frequently persists as a moderate amplitude feature. On the greater number of such occasions, the next downstream trough is of large to moderate amplitude toward the end of the four day period.

(iii) The majority of existing troughs within the broad sector eastwards of a newly formed primary either decrease in amplitude or become degenerate during the following four day period. The large amplitude downstream troughs however, though present in relatively small numbers at all times, show little change in frequency during the period.

(iv) Some new troughs develop within the sector and period considered but those with small and medium amplitude are insufficient to make good the loss of existing features referred to in (iii). The formation of new large amplitude troughs is infrequent but shows a slight tendency to increase towards the end of the period.

/(v)

by their approximate longitudes.

Simple Growth.

January 19th, 1950. — Trough of amplitude (A) 5°E.
This trough formed by simple growth of a trough in 5°E on January 17th.

February 15th, 1950. (Fig.2). — Trough amplitude (A) 40°W,
A vortex was present on February 12th in 70°W; the associated trough, (and ridge immediately to E) slowly increased in amplitude.

August 24th, 1950. — Trough amplitude (A) 10°W.
Temporary increase in amplitude of a persistent feature.

November 6th, 1950. — Troughs amplitude (A) 15°W and 10°E.
Both large amplitude troughs resulted from temporary increase in amplitude of pre-existing features largely due to intensification of the intervening ridge.

November 8th, 1950. — Trough amplitude (A) in low latitudes 25°W.
The vortex in 15°W on November 6th moved steadily S. and intensified S. of main belt of westerlies.

Amalgamation or "Phasing in".

March 8th, 1950. — Trough amplitude (A) 45°E.
The trough in 30°E on March 5th, phased in with the trough at 50°E on March 5th, the resultant trough subsequently absorbing the slow moving vortex in the Mediterranean.

(v) Combining (iii) and (iv) we conclude that the most usual course of events following the formation of a primary trough over America is for the downstream pattern to become complex or of small amplitude. There is however an important minority of occasions (associated largely with the persistent primaries), for which the incidence of large amplitude downstream troughs increases during the period to reach a maximum frequency of about 30% for contours and 45% for thickness troughs on the 4th day when troughs of all periods of persistence are considered.

For the reasons given in paragraph 2 the above conclusions apply only to those cases in which the primary troughs are of "simple" structure and with axes lying nearly in the north-south direction. It is hoped that they have a broad application to more complex primary troughs although the present report offers no evidence on this point.

References.

Eady, E.T. 1950. Centenary Proc. R.Met.Soc., London.

Rossby, C.G. 1945. Journal of Met., Lancaster, Pa., Vol.2, No.4.

1949. Tellus, Stockholm, Vol.1, No.1.

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No. of cases (Primary B)	0	0	1	3	0	1	0	0	2	4	11
Total	0	2	3	4	3	2	2	1	2	14	33

The scatter is very large indeed. Since the mean westerly zonal current at 500 mb. over the period and half-hemisphere varied between 5° and 20° per day it is clear that the majority of the new troughs were not the result of simple advection. Rossby (1949) suggests that the maximum dispersive increment is about 18° per day but, allowing for this and extending the propagation speed to 40 degrees per day, only half the occasions are accommodated. From this point of view about half the occurrences might be dispersion effects (but of course are not necessarily such) and about half almost certainly are not.

Our data do not therefore lend much support to the dispersion hypothesis and certainly do not encourage the practical use in forecasting of formulae based upon it. This does not mean however that downstream effects do not occur, certainly on occasions the development of one oscillation is clearly linked with an earlier disturbance elsewhere, but the mechanism of the linkage probably involves the processes of baroclinic development.

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APPENDIX

Mode of formation of large amplitude downstream troughs.

The data for the year 1950 showed 22 distinct large amplitude (A) downstream troughs associated in a broad sector to the east with primary troughs of amplitude (A) or (B), over the period D0 - D4.

Of these, six could be attributed to simple growth or increase in amplitude of existing troughs. Most of these cases however involved a change of amplitude from (B) to (A) for a period of the order of one day only, the change of amplitude not being really significant.

In 11 cases, the formation of a large amplitude trough arose through the amalgamation or "phasing in" of troughs, frequently of different latitudes. The commonest of these processes appeared to be due to a trough in a mobile wave train north of the main belt of westerlies, (often initiated by the shearing off of the northern portion of a large amplitude trough with consequent new trough formation), phasing in with a trough in the main westerlies.

The five remaining cases could be associated with "blocking" patterns; either with a retrogression and intensification of the trough to the east of the blocking complex, or with the breakdown of the block and the absorption of the blocking low by the trough to the east or west of the complex.

From this study it would appear that large amplitude (A) downstream troughs are perhaps more likely to result from the interaction of wave trains of different latitudes than from any other process. The large troughs developing in association with blocking patterns may frequently be regarded as special cases of this process.

Details are given below of the occasion and mode of formation of the large amplitude downstream troughs studied. Figures 2, 3, 4 and 5 show examples of each of the processes mentioned above. The locations of the troughs are given by their approximate longitudes.

Simple Growth.

January 19th, 1950.	Trough of amplitude (A) 5°E. This trough formed by simple growth of a trough in 5°E on January 17th.
February 15th, 1950. (Fig.2).	Trough amplitude (A) 40°W, A vortex was present on February 12th in 70°W; the associated trough, (and ridge immediately to E) slowly increased in amplitude.
August 24th, 1950.	Trough amplitude (A) 10°W. Temporary increase in amplitude of a persistent feature.
November 6th, 1950.	Troughs amplitude (A) 15°W and 10°E. Both large amplitude troughs resulted from temporary increase in amplitude of pre-existing features largely due to intensification of the intervening ridge.
November 8th, 1950.	Trough amplitude (A) in low latitudes 25°W. The vortex in 15°W on November 6th moved steadily S. and intensified S. of main belt of westerlies.

Amalgamation or "Phasing in".

March 8th, 1950.	Trough amplitude (A) 45°E. The trough in 30°E on March 5th, phased in with the trough at 50°E on March 5th, the resultant trough subsequently absorbing the slow moving vortex in the Mediterranean.

April 11th, 1950. Trough amplitude (A) 5°E.
A temporary increase in amplitude of a pre-existing large amplitude feature, perhaps explained by the absorption of the slow moving vortex in the E. Mediterranean on April 9th.

April 19th, 1950. Trough amplitude (A) 5°E.
The trough in 35°W on April 19th phased in with the next eastward trough.
The American member of the trough pair spanning the Atlantic on April 17th moved E. to give good ridge and amplitude (A) downstream trough on April 19th.

April 23rd, 1950.(Fig.3.) Trough amplitude (A) 5°E.
The northern portion of the tilted trough in 30°W. on April 21st, ran eastward to phase in with the next eastward trough.

May 5th, 1950. Trough amplitude (A) 50°W.
The trough formed by amalgamation of a vortex in 60°W. and the trough pair SE. and SW. in 85° and 40°W. on May 2nd.

May 9th, 1950. Trough amplitude (A) 60°W.
A trough in 75°W. on May 7th, took on a more simple structure with the absorption of the vortex to the SE. at 40°W.

August 15th, 1950. Trough amplitude (A) 20°W.
The northern portion of the trough in 60°W. on August 13th, ran ahead and subsequently extended towards the pre-existing trough in the Mediterranean.

October 16th, 1950. Trough amplitude (A) 15°W.
The trough in 20°W. on October 13th, phased in with the vortex developing in low latitudes in 5°W. on October 14th.

December 2nd, 1950. Trough amplitude (A) 00°W.
A member of a family of vortices around the Pole on November 28th, at 70°W. on this date, moved rapidly E. to the Iceland region where it intensified on engaging the flow associated with the main westerlies.

December 11th, 1950. Trough amplitude (A) 5°W.
The northern portion of the trough in 40°W. on December 9th, moved ahead of the southern portion to phase in with the next eastward trough.

December 22nd, 1950. Trough amplitude (A) 5°E.
The trough associated with a small closed vortex in 50°W. on December 18th, ran E. to amalgamate with the next eastward trough.

Blocking Patterns.

January 28th, 1950.(Fig.4.) Trough amplitude (A) 45°E.
The blocking pattern of January 24th was disturbed by a baroclinic feature running round N. of the block and developing on its E. side, effectively retrograding the pre-existing tilted trough.

March 29th, 1950. Trough amplitude (A) 40°E.
This feature soon became complex. The trough in 25°W. on March 25th cut off a vortex in the south and the upper portion continued E. and developed,

the trough plunging S. to the E. of the blocking pattern established on March 27th.

May 15th, 1950. Trough amplitude (A) 20°E.
The trough in 10°W. passing to N. of blocking pattern on May 13th, intensified and effectively caused block to retrogress.

May 19th, 1950. (Fig.5a) Trough amplitude (A), 10°W.
The blocking pattern of May 17th broke down and the blocking vortex west of Spain was absorbed by the trough over the British Isles and Scandinavia.

November 27th, 1950.(Fig.5b) Trough amplitude (A) 30°W.
A trough formed with breakdown of the blocking pattern in mid Atlantic on November 25th. The trough in 50°W. phased in with the low-latitude vortex of the blocking complex.

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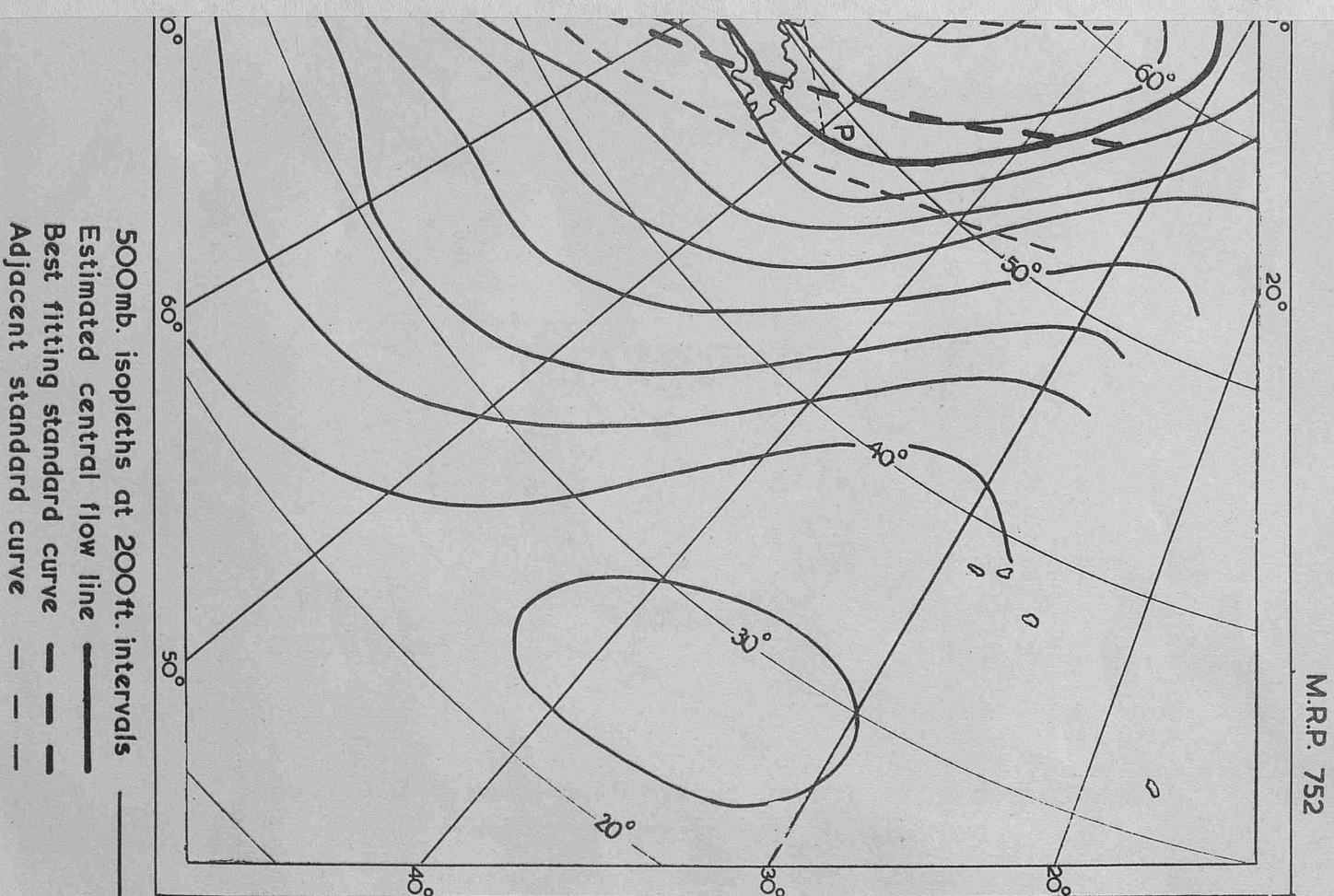


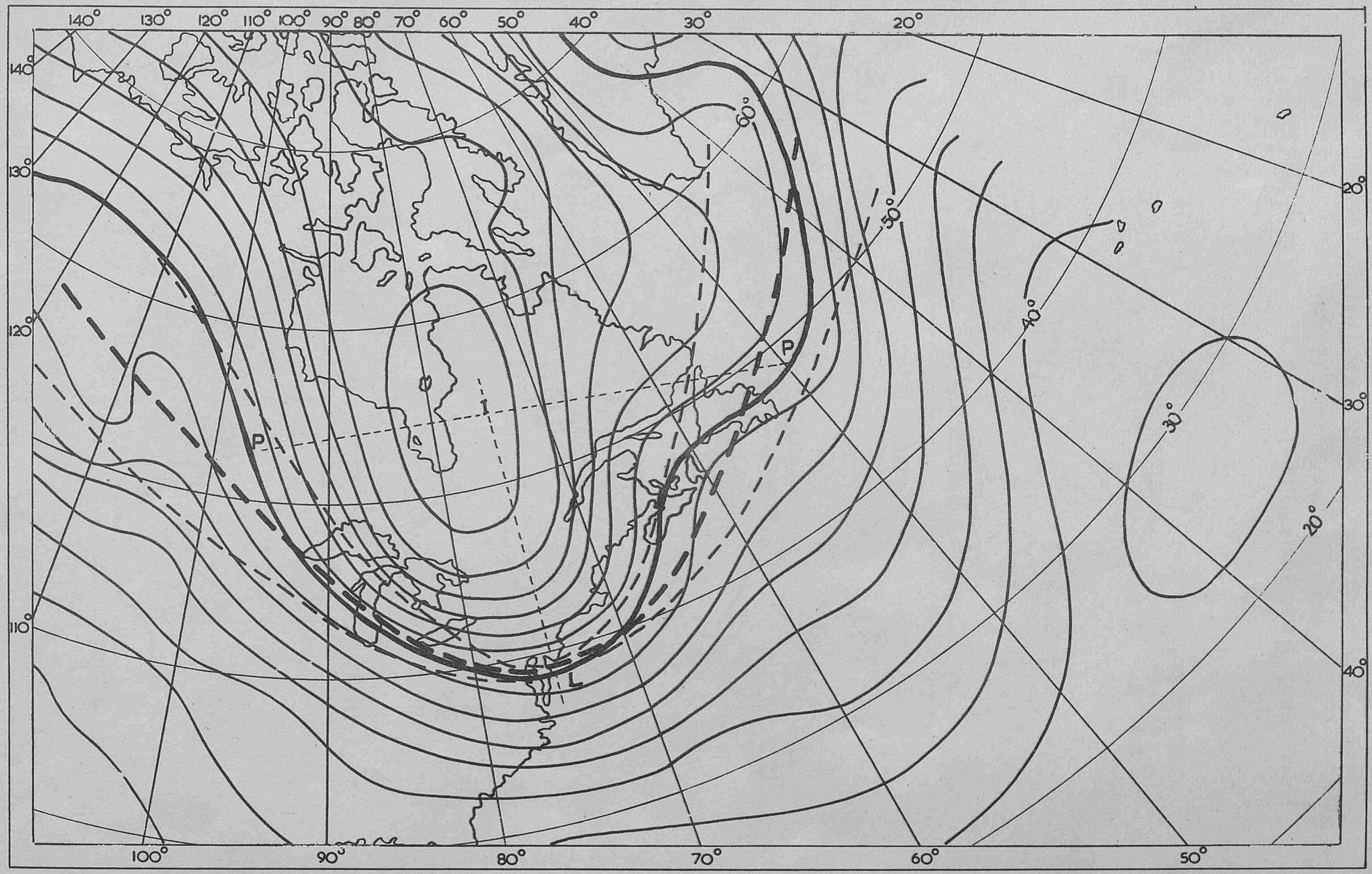
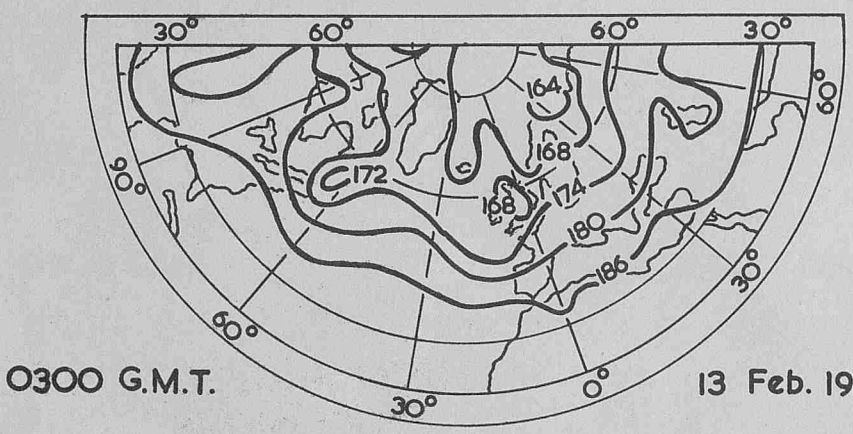


Fig.1 Example of simple primary trough.

500mb. isopleths at 200ft. intervals ———
Estimated central flow line ———
Best fitting standard curve — — —
Adjacent standard curve — — —



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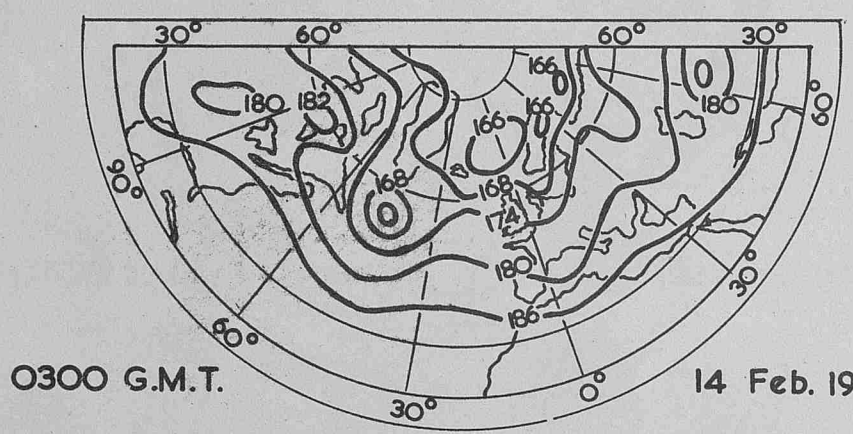

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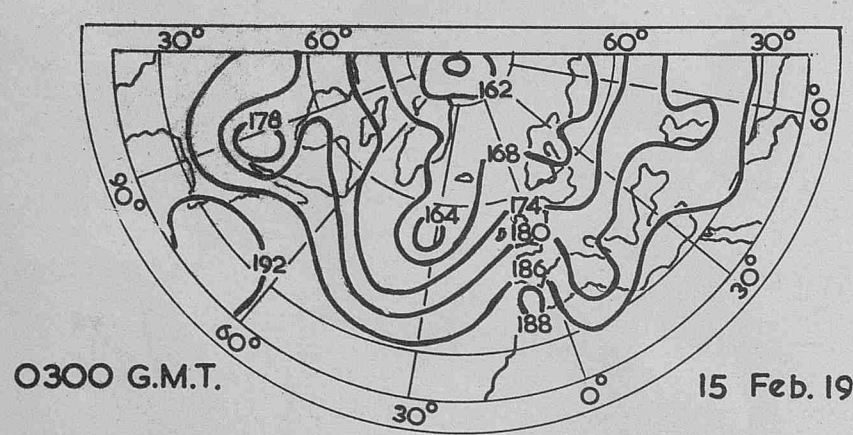


0300 G.M.T. 15 Feb. 1950

Fig. 2 Example of simple growth of a downstream trough. Primary over N. America; downstream trough over Atlantic.

500 mb. contours ——— Heights in hundreds of feet.

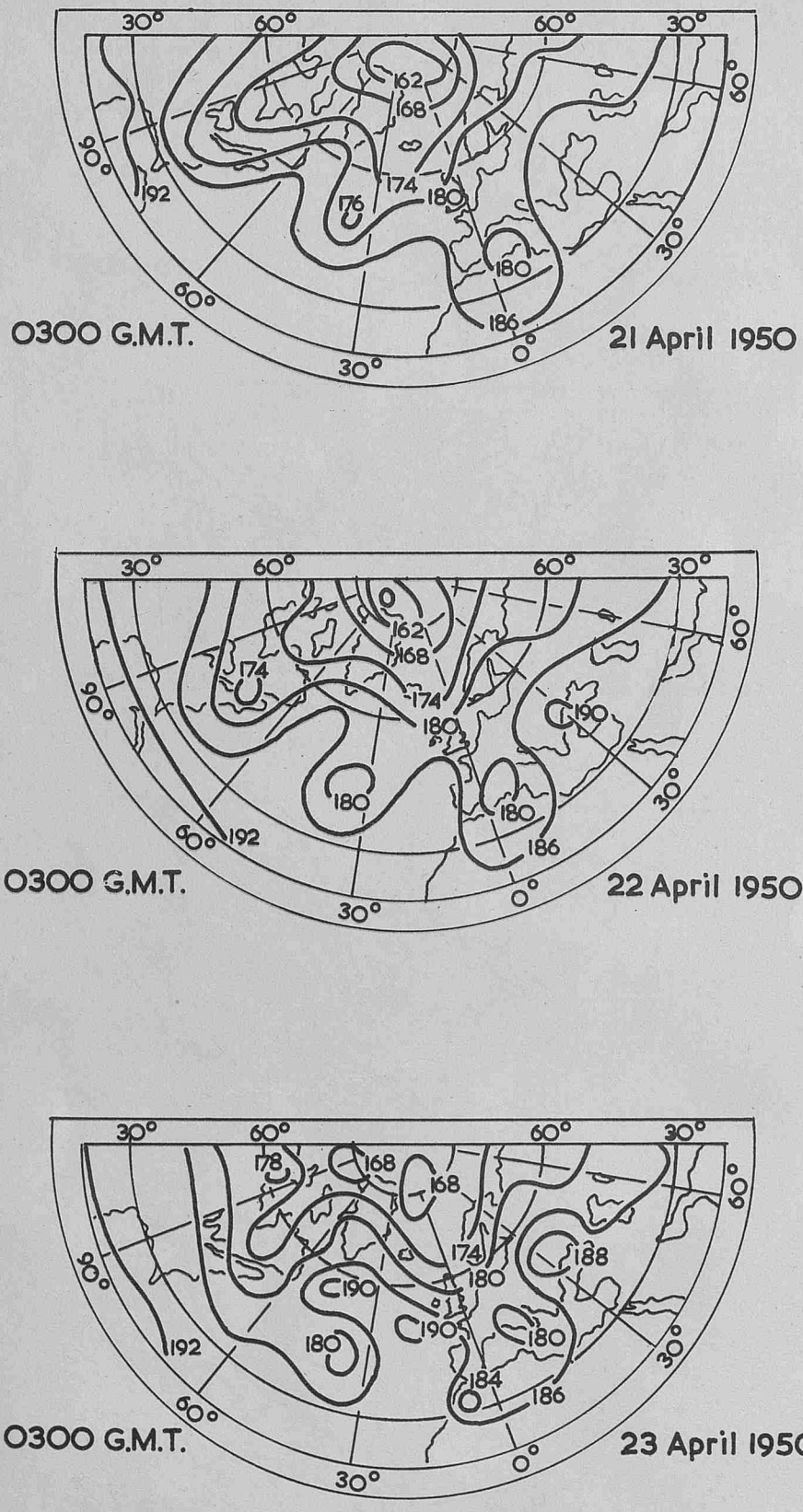
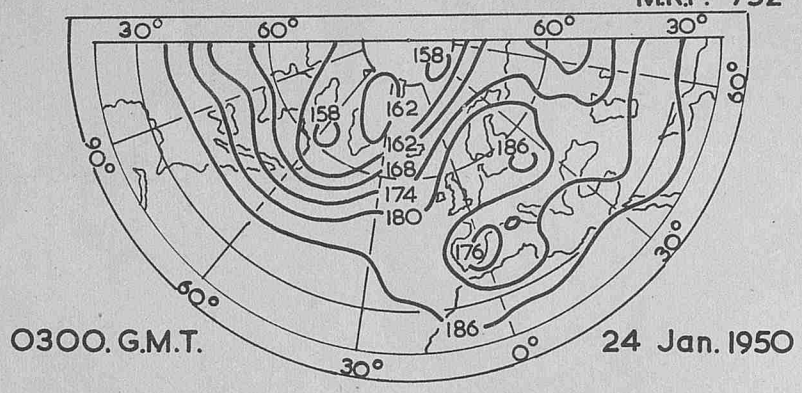
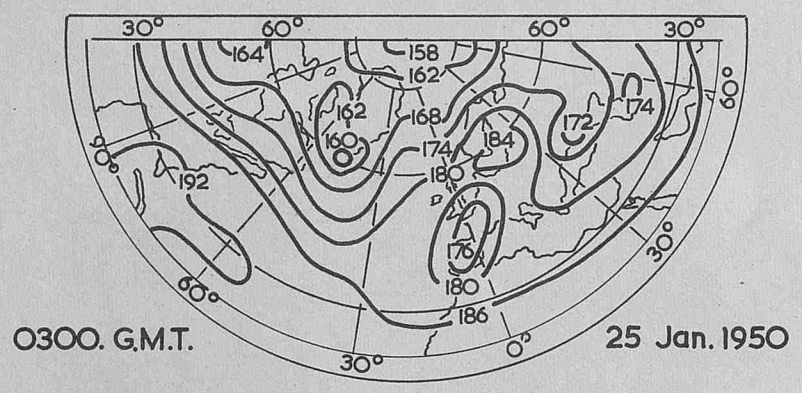


Fig. 3 Formation of large amplitude trough by the phasing in of mobile trough in high latitudes with stationary trough over Mediterranean.

500 mb contours ——— Heights in hundreds of feet.



0300. G.M.T. 24 Jan. 1950



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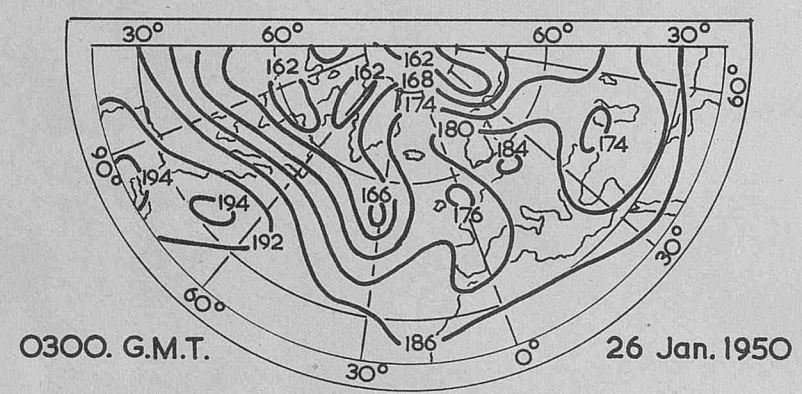

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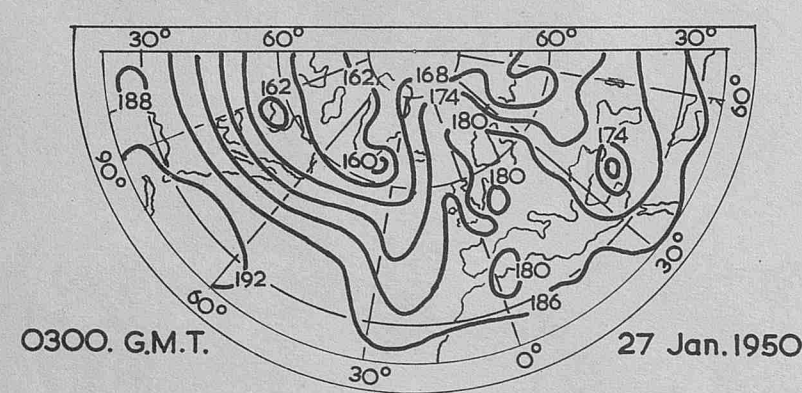

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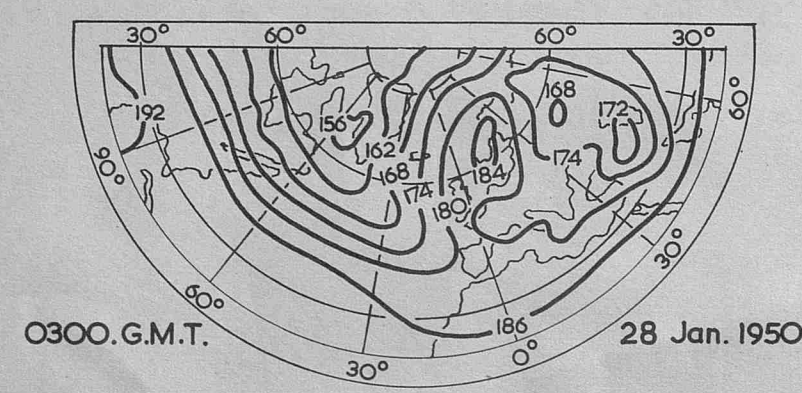
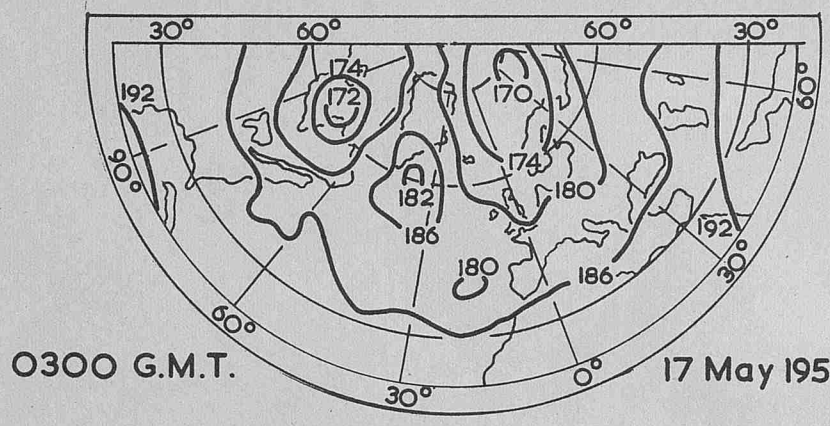
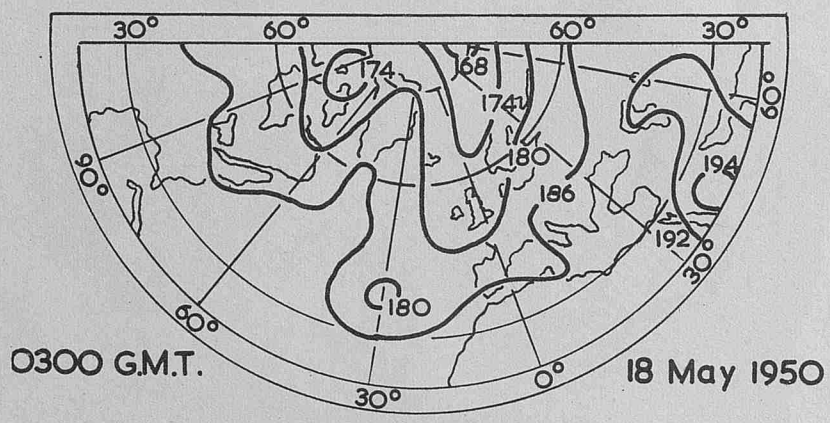


0300. G.M.T. 28 Jan. 1950

Fig. 4. Development of large amplitude trough over Russia in association with blocking pattern over Western Europe.

500mb. contours ——— Heights in hundreds of feet.





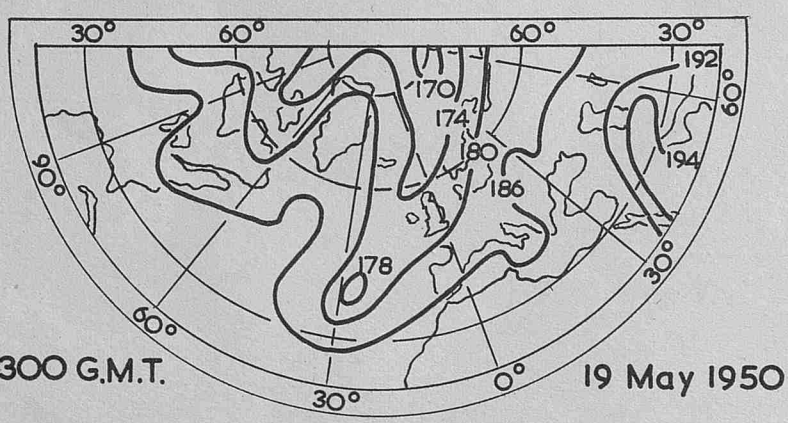
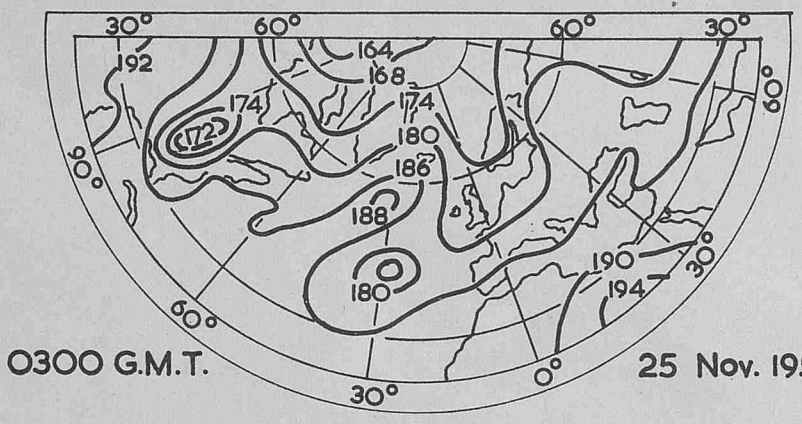
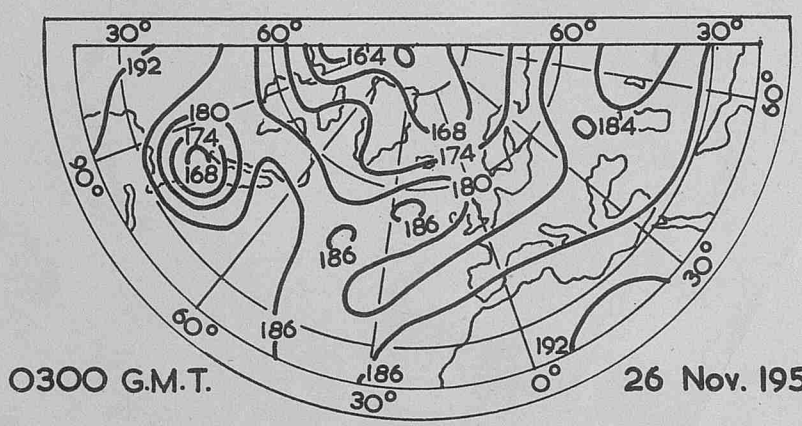


Fig. 5(a). Example of large amplitude trough resulting from breakdown of blocking pattern.

500 mb. contours ——— Heights in hundreds of feet.





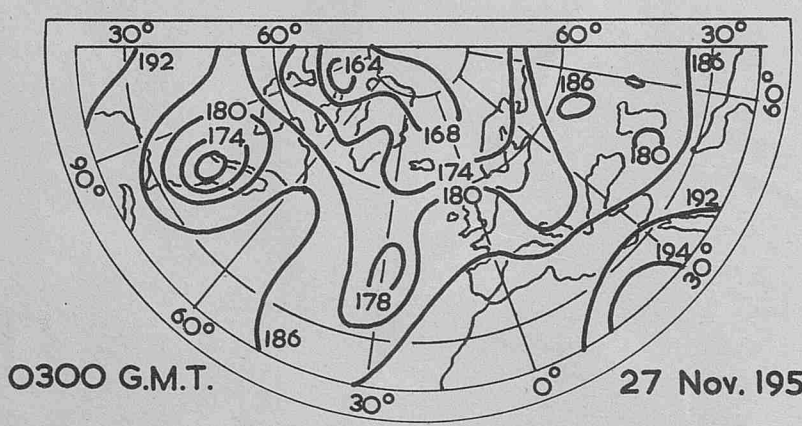


Fig. 5 (b). Example of large amplitude trough resulting from breakdown of blocking pattern.

500 mb. contours ——— Heights in hundreds of feet.

M.R.P. 755
S.C.II/120.
20th September, 1952.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No ..... by Mr. ..... A copy is available in the Library of the Meteorological Office."

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THE FORMATION OF NEW ANTICYCLONES

by R.C. Sutcliffe (Meteorological Office)

ABSTRACT.

Defining a new anticyclone simply as the emergence of a centre of relatively high pressure on the surface chart not continuous with any pre-existing centre (but omitting weak short-lived cells in a ridge situation) some 42 cases were found in the two years 1950-51 over the area of western Europe, eastern and central Atlantic. These are classified according to 500 mb contour and 1000 - 500 mb thickness patterns. Only four cases were associated with simple oscillations in a baroclinic zone with a pre-existing "sub-tropical" anticyclone to the south. 21 cases were also simple oscillations but with low surface pressure pre-existing to the south. Nine cases were associated with upper "anticyclonic wave disruption" and the cutting off of a low latitude cold pool, the high developing across the neck of the cut-off with surface cyclogenesis further south. Eight cases were complex.

Qualitatively the associated upper patterns were much more frequent than the new anticyclonic cells although anticyclonic building was typical. It is concluded that anticyclonic development in this region is normally shown by the formation of a ridge or by the movement or intensification of an existing high and that the appearance of a new high cell is a matter of quantitative refinement rather than a distinguishable dynamical type. In practical forecasting the patterns will indicate anticyclogenesis but whether or not to introduce a new cell must, in the absence of quantitative dynamical methods, rest on synoptic judgment. The criterion of new anticyclone formation as adopted is not a suitable one for the study of anticyclogenetic problems in general, although the statistics may be a useful contribution to synoptic climatology.

1. INTRODUCTION

The formation of new depressions and new anticyclones has always presented a peculiar difficulty in forecasting. This is naturally so when traditional methods of extrapolation of the pressure field are used, for one cannot extrapolate what does not exist, and attempts to anticipate new pressure features from higher order derivatives (say the variations in the field of tendency - isallobars) although helping a little do not get very far. The better dynamical understanding which we now have helps a little, in so far as we can recognise upper-air situations which are liable to give rise to new

/developments

developments, but also suggests that extrapolation, whether empirical or dynamical, must have a limited scope. The new systems, we now understand, are often instability developments (in the sense that they emerge by self-development processes from small beginnings,) and, taking into account the inadequacy and errors of our observational data which mask the earliest stages, the new developments appear effectively as acts rather of creation than of evolution. Nevertheless, we are broadly aware of the synoptic models of cyclogenesis and anticyclogenesis and the object of the present investigation was to discover whether those occasions when a new centre of high pressure emerges can be usefully classified according to the associated upper level patterns.

The only possible definition of an anticyclone, consistent with synoptic usage, is a closed centre of relatively high pressure on the surface chart and with this definition it is possible to pick out the occasions of a new formation by inspection of the surface charts alone. The only criterion is that a centre should appear over the area of study (taken broadly as Western Europe and the Atlantic excluding the far west and shown in Figure 1) and should not be continuous with a pre-existing centre either inside or outside the area. It was realised and accepted that a sample of cases chosen in this way would not be a general representative sample of the dynamical process of anticyclonic development defined perhaps by the development of anticyclonic vorticity at low levels, perhaps by marked subsidence, perhaps by an appropriate field of isallobars. Such a wider study might be undertaken but it would open the flood gates to every ridge development and every case of building or displacement of an existing anticyclone and would have little connection with the problem of the new anticyclone.

The difficulty anticipated and encountered lies in the fact that the normal effect of a new anticyclogenesis in our area is not a new anticyclone, as defined, but a mere ridge extension from, or movement and development of, an existing high. In order to exclude the trivial cases where a closed isobar was to be found for a time in what was essentially a ridge structure a case was admitted only when the cell appeared on the charts of the Daily Weather Report and persisted for 24 hours at least. This restriction eliminated a number of small scale, weak centres of local importance but was necessary if we were not to be in difficulties with minor details of chart analysis. It was not however thought proper to include any of the numerous cases when even vigorous anticyclonic building took place without the creation of a distinct new centre since there could be no criterion of what to include if the new centre were non-essential.

The sample has the virtue of being selected tolerably objectively from the surface charts alone, so permitting the classification to be made subsequently by reference to the upper air charts. Over the two years 1950 - 51 a total of 42 cases was obtained in this way. The 500 mb contour charts and the 1000 - 500 mb thickness charts are taken as the basis for classification.

2. THE CLASSIFICATION

Although the classification into types of new anticyclone formation which is arrived at may seem obvious enough as it is presented it was not so in the investigation. Synoptic situations are infinitely varied and to pick out common features of situations represented each by three series of charts over a period of days is not easy unless one has a definite thesis in mind. However, in due course it was recognised that in most cases the new surface centre was associated with simultaneous and characteristic variations in the upper patterns thus establishing the fact that we were dealing with three-dimensional dynamical developments which could be studied to some extent by the technique. There were some 20% of cases of much complexity when the local patterns were confused by simultaneous and more or less independent developments in the vicinity but no case was found where the development viewed after the event seemed synoptically freakish or mysterious.

/With

With reasonable freedom in interpretation, every case, with the few "complex" exceptions, could be fitted into two models related with the upper patterns, that is with the wave-like structure of the upper contours. These two models may be called the simple sinusoidal oscillation and the wave disruption: they are well known but are explained in more detail.

3. THE SIMPLE SINUSOIDAL OSCILLATION

The growing (unstable) wave-model of the depression in a baroclinic zone is generally familiar, theoretically well-founded and synoptically common. The corresponding anticyclone model is perhaps less well known but the mechanism has been discussed briefly by Sutcliffe and Forsdyke (1950). There is still some doubt as to the exact criteria of "instability" if account is taken of all the controls but certainly something of the kind is present somewhere at all times and perhaps every steady baroclinic current is unstable for some such development of appropriate synoptic scale.*

3.1 The classical general circulation model

In the classical model of the general circulation of the atmosphere, with the unsettled westerlies lying on the poleward side of the sub-tropical anticyclone belts, the typical sequence is one of developing, occluding and ultimately filling depressions separated not by closed anticyclones but by more or less intense wedges or ridges extending from the semi-permanent highs. Although this model is accepted and does not lack dynamical explanation the rarity with which a new high centre was in fact formed when the broad weather type in the region fitted this general circulation model came as something of a surprise. The broad weather type is of course common over the Atlantic sector with the Azores anticyclone setting the pattern and the degree of development of the baroclinic ridge is a regular and important problem for forecasting. But, excluding the cases where there was a temporary and weak closed centre, usually behind the cold front, but where otherwise the ridge construction was satisfactory, it was difficult to find a clear case of new anticyclone formation associated with the simple upper oscillation. Typically the ridge would subside within a day or two and even in the cases where the development was strong and led to a large change in the broad weather pattern it was manifest by the old warm anticyclones being displaced to higher latitudes rather than by a definitely new surface formation. In Fig.2 is illustrated a good example of pronounced anticyclonic building of this type. A few comments on the case follow.

Period 19th - 21st October 1950 - Figure 2

On the 19th there are two developing depressions in the Atlantic in the heart of the baroclinic zone shown by the thickness lines: the 500 mb chart shows a rather weak sinusoidal pattern. The lower latitudes are dominated by the "sub-tropical" anticyclone lying southwest of the British Isles and outside the main upper stream of westerlies. The two depressions are separated by a "baroclinic ridge" from a detached high cell to the south. Over the next two days the depressions move east and the amplitude of the upper wave pattern grows considerably, the surface ridge develops at the same time and by the 21st has effectively absorbed and displaced the old warm

/anticyclone

*Footnote. While the author is ready to accept the idea of perpetual instability, propounded for example by Eady (1949), as compatible with synoptic experience he feels that the theory must further show why we get a finite number of synoptic systems on the synoptic scale giving coherent patterns, not chaos. Presumably it is due to the fact that the flow once organised into a developing sub-system is no longer unstable for other developments of similar scale in the same region but only for still smaller scale "secondaries" or turbulence. Perhaps this is an elementary matter of theory but it does not appear to have been specifically treated.

anticyclone into the baroclinic zone. There is, strictly speaking, no new anticyclone formation; rather the case illustrates the merging of two high centres into one. But there is very obviously a pronounced process of anticyclogenesis of the simple oscillation type. The case, apart from being a nice example of the process, also illustrates how the criterion, "the formation of a new high centre", must fail to pick out the cases of anticyclogenesis when the development takes place north of a pre-existing warm anticyclone corresponding with the classical general circulation organisation. It is clearly a trivial matter whether there are two centres, as on the 20th, or only one, as on the 21st; a similar evolution often occurs with no separate centre north of the cold front at any stage.

In the two years only four cases of this type were picked out showing a distinct new surface centre and none was very clear-cut or without some complicating factor. The conclusion is that anticyclogenesis associated with a simple oscillation in the upper westerlies north of a high pressure belt may lead to a ridge (perhaps containing a weak separate cell) which is absorbed by the old sub-tropical high or to one which itself absorbs the old high but very rarely to a distinct and detached new centre. Although the ridge model is well known to be typical the rarity of the new detached high centre is rather remarkable when one recalls that a new cyclone centre regularly appears in the quite early stages of almost any cyclonic oscillation.

3.2. Non-classical broad-scale situation

The position is very different when the broad weather situation is not roughly that of the classical general circulation model, when the upper baroclinic westerly in which ridges and cyclones develop does not lie poleward of an established high but has generally low surface pressure at lower latitudes. This pattern can arise in many ways but there are certain rather common cases in our region. Two examples are illustrated in Figures 3 and 4.

Period 9th - 11th June 1950 - Figure 3

In this case the zone of upper westerlies as shown on the 500 mb charts is north of its mean position and, as in the period 19th - 21st October (Fig.2) we have two baroclinic depressions separated by a ridge extending, in this case, from the large high away to the southwest. There is however, on the 9th, generally low pressure over southwest Europe associated with a cut-off cold upper air vortex. As the sinusoidal wave pattern in the westerlies moves east and grows in amplitude the ridge breaks off from the high to the west and we have a clear case of a new anticyclone. The example shows how a simple oscillation in the westerlies may readily give rise to a new high cell when there is no pre-existing anticyclone to the south of the anticyclogenetic region of the wave pattern and a ridge construction is then hardly possible on geometry alone.

Period 30th November - 2nd December 1950 - Figure 4

There is clearly much similarity, in a broad sense, between this and the previous case. The main westerlies and the main baroclinic zone are again well north and there is already a large amplitude oscillation. The low latitude surface depression in mid-Atlantic is completely cut off from the upper westerlies and is also thermally cut off as a cold pool. With further growth in the amplitude of the eastward-moving upper wave pattern a new high centre is formed on the 1st, this time appearing to detach itself from the high latitude anticyclone. One can well imagine that if a similar oscillation were to arise in the absence of the low latitude depression (that is if the Azores high were present as in the classical general circulation model) there might well be only a ridge development with no new high cell. Incidentally, the Atlantic situation on the 30th may be regarded as a "block" (higher latitude high and lower latitude low with separation of the upper

/westerlies

westerlies into two branches) and we see a tendency for the block to be resolved on the 2nd by the cut-off, low-latitude, cyclonic vortex at 500 mb being swept up, as it were, into the trough extension from the northeast.

The last two examples show clear cases when the new high, associated with a large oscillation of the westerlies, became a major feature near the British Isles; but, quite generally, whenever the upper-westerlies run north of a low pressure region high cells are readily formed even with quite small scale oscillations. Thus it is common to have a more or less normal sub-tropical Atlantic anticyclone with a col extension across southwest Europe north of low pressure in the Mediterranean (common in winter) or over North Africa (typical of summer). Then, associated with ridges between travelling higher latitude depressions, new high cells appear to detach themselves from the Azores high and move eastwards across Europe. These cells may be quite important for detailed, short-period, forecasting. There were several cases in July, 1950.

Something rather similar occurs when there is an occluded cold depression, say over the British Isles, with high level westerlies continuing in a baroclinic zone at rather high latitudes (analysed as containing the Arctic front). Then new high cells may appear to detach themselves from a semi-permanent high in the polar seas and move east or southeast into Scandinavia. These occur usually with quite conventional oscillations in the upper westerlies of these latitudes. There were two cases in November 1951.

The ease with which new high centres form when the broader pre-existing pattern favours the new cell rather than the ridge structure suggests that the relative infrequency of new highs over our region is due to the normal large-scale general circulation pattern and may be considered in relation with the further fact that the classical general circulation model is not at all typical of the continents. Although no statistical study of these regions has been made in this context it is the impression that new high centres form readily over North America in the baroclinic zone and not only in the winter season when a "thermal" explanation might be offered. Over Eurasia too, the mobile high cell is a normal feature, a fact which no doubt accounts for the emphasis given to the anticyclone in the Russian system of forecasting. One recalls that the Multanovsky technique is concerned largely with anticyclone tracks, that is with mobile baroclinic anticyclones which are typical of the Continental area. It must not be too hastily assumed that a similar prominence could properly be given to anticyclonic tracks in our region: the general circulation background is very different.

4. THE DISRUPTING OSCILLATION

For the most part (but rarely over the whole hemisphere at the same time) the picture given by a study of the upper westerlies of middle latitudes and the associated surface pressure features is of one baroclinic zone with a number of wave-like distortions passing between cold troughs and warm ridges these being linked with the surface features according to various models which it is not necessary to describe here. New cyclonic centres regularly, and new anticyclonic centres in our region less frequently, are associated with new sinusoidal oscillations in the upper current but the main upper stream remains concentrated around one maximum. From time to time however, especially when the oscillation reaches a large amplitude, the current and the wave pattern disrupt, the stream divides and two distinct wave-like distortions move at different speeds rapidly becoming out-of-phase. This process of disruption has been much described in connection with "blocking" (see e.g. Rex 1950, 1951) and "cutting-off" (Palmén, 1949) although no satisfactory explanation of the dynamics or criterion for the occurrence is yet available. (Rossby (1950) has offered a relevant suggestion but it does not seem quite to fit the facts).

/The

The problem of disruption is likely to be difficult until at least we have an acceptable quantitative explanation of the spatial coherence of the upper wave pattern and the associated depressions and anticyclones which do persist as features in spite of the very large variations in wind speed both horizontally and vertically. It is perhaps more remarkable that any large scale wave-like form can hold together than that it should disrupt from time to time. It is however probably relevant to note that a wave-form in the westerlies if moving with uniform speed over the earth's surface will tilt forward at high latitudes merely owing to the figure of the earth while the Rossby retrogressive effect on the phase velocity $\beta L^2/4\pi^2$ (where β is proportional to cosine of latitude and L is the wave length) almost certainly has a direct bearing on the relative stagnation of the lower latitude wave-form once disruption has occurred and may be an important factor in the dynamics of disruption.

While geometrically we may imagine the wave-form disrupting by relative progression either at the higher or the lower latitude the former would on the above considerations seem the more likely and in fact it does occur much more definitely and frequently. A schematic representation of the sequence of charts is shown in Figure 5. It is characterised by the lagging behind of the low latitude upper trough followed by complete disruption and the cutting off of a low latitude upper low generally with a cold pool and, in our region, also a low latitude surface depression; at the same time high pressure builds across the neck of the cut-off.

As the relative shear between the two parts of the trough is anticyclonic in sense, with the surface ridge across the neck appearing rather like a roller, the process will be called "anticyclonic disruption". Some nine of our new anticyclone centres appeared in this way one of which is illustrated in Figure 6 discussed below. It must however be noted that the process does not normally produce a new high cell, according to our definition, but more commonly gives only a ridge development or a displacement across the neck of a pre-existing high from the west. Thus in 1950 there were some 33 cases of anticyclonic upper disruption over the region; all gave a ridge building across the neck and all gave a cut-off low latitude depression but only five clearly satisfied our criterion of formation of a new high centre: as in the case of simple oscillation the anticyclogenetic process is much more common than the emergence of a new, detached, high centre.

Period 25th - 27th March, 1950 - Figure 6

On the 25th the 500 mb westerlies show one broad current across the Atlantic although the large-amplitude trough is already broken across the neck. At the surface there are two depressions associated with this upper trough, one to the east of the southern limit of the trough, the other further north. By the 26th the upper disruption is complete, the high latitude part of the trough has moved to the north of Scotland lagging in phase behind the associated surface low in the normal way. The low latitude part of the upper trough has moved more slowly and become a well separated vortex almost in phase with the cold pool and the surface depression. A new anticyclone now detaches itself from the ridge to the west and builds "across the neck". By the 27th higher latitude progression has continued, the new anticyclone is phased into the development region between the two depressions and is due north of the low latitude cut-off low (so constituting a block in the westerlies).

Although anticyclonic disruption is the common way in which a simple upper wave-pattern may disrupt and the stream divide an inspection of the upper charts alone will reveal cases where the southern part of an upper-trough breaks forward and undercuts a higher latitude ridge: a cold cut-off lower latitude depression is the normal outcome. The process is found to westward of a blocked situation and is well illustrated by Berggen, Bolin and Rossby (1949). The immediate controlling system is a baroclinic cyclogenesis in the westerlies as they approach a strong difluence and may be called

/a

a "cyclonic disruption". Although it is associated with a blocking high, even with its development or retrogression, there was no clear case of a new anticyclone appearing in the process over the two years considered and, in spite of the importance of the situation in the general problem of anti-cyclones, it calls for no further attention here.

5. COMPLEX SITUATIONS

Although it has proved possible to classify some 80% of the cases into the two types of simple oscillation and wave-disruption there remain a number of cases which can only be called complex, where more than one region of dynamical development, more than one system of depressions, although to some extent independent, interact giving rise to a new high cell as an element in the complex. This sort of process is to be expected and it is perhaps surprising only in that the number of cases is so small. The only approach likely to lead to useful comment is by quantitative dynamical analysis of each occurrence and in the present state of quantitative methods one would not choose the unusually complicated as an exercise. One example is however illustrated.

Period 20th - 22nd February, 1950 - Figure 7

This is a clear case of two baroclinic zones over the Atlantic with two sets of frontal depressions. Between the two a new high cell is detached on the 21st in mid-Atlantic and this later divides, one part phasing in between the pair of depressions in the north, the other dropping into place appropriately in the southern system over Europe. The northern high may be regarded as developing with the growing 500 mb oscillation, as a simple case of the class studied in Section 3.2. But between 20th and 21st we have deepening of four neighbouring baroclinic depressions simultaneously, a situation which is extremely rare and may well be unprecedented in recorded meteorology.

6. SUMMARY AND CONCLUSIONS

All the cases of new anticyclone formation, over the defined area and the two-year period, recognised by the formation of a new centre of high pressure on the surface chart (and excluding weak cells in a basic ridge pattern), were, with a number of "complex" exceptions, associated either with a simple oscillation in the upper baroclinic current or with an "anticyclonic disruption" of the upper wave-like pattern. The list of occasions, so classified, is given in Table I.

The associated upper-air patterns, defined in this qualitative way, are however much more common than the occurrence of a new high centre. They are indeed the patterns which normally occur with anticyclonic dynamical development generally. Thus the upper oscillation is associated with any series of depressions but whereas in the two years only 25 cases were picked out the number of ridge developments is too numerous to be worth the counting. Again, the anticyclonic disruption occurred 33 times in one year, always with a surface building of high pressure across the neck of the disruption, but only five cases showed a distinct new high.

The conclusion is that the study has failed to provide any useful criterion for the new formation which emerges only as a rather exceptional variation on anticyclogenetic processes which do not normally break the geometrical continuity of pre-existing high pressure centres. It would be reasonable to look for some extra criterion to separate the exceptional event from the more usual but the implication of the evidence and known theory is that any such criterion would need to be expressed as a quantitative relationship between all the factors which are involved in surface pressure changes and would not be found in any distinct and qualitatively recognisable

/characteristic

characteristic of the pre-existing patterns. In other words one would expect the new centre to be predicted only if the dynamical and thermodynamical problem could be solved with accuracy whereas, in the present state of theory, it is as much as we can do to infer anticyclonic building qualitatively without the refinement of predicting that a new centre must emerge rather than a ridge extension from or a development of a pre-existing neighbouring centre. Attempts are being made, in this country and elsewhere, to develop quantitative techniques and we may have to await the outcome of this work before we are able to judge the practicability of forecasting the new high cell.

Meanwhile we may conclude that the formation of a new high centre is to be looked for whenever the upper air models of anticyclonic development chance to occur with surface patterns which render the ridge or displacement construction geometrically awkward: and we should perhaps not be afraid to make the prediction in these cases merely because the event is statistically infrequent. Obviously, also, the awareness of the possibility will help to give confidence in the early extrapolation of isallobaric fields.

Nevertheless, although the problem is probably one of "dynamical instability" there remain aspects which synoptic study may yet clarify. It may not be particularly important that a new high cell should form instead of a mere ridge or displacement and intensification of an existing centre but it is all important to know whether a new anticyclogenesis will have a major effect on the large-scale patterns or give merely a temporary small-scale feature.

The theoretical study of unstable wave perturbations in a baroclinic current leads, in the first analysis, to the development of alternating depressions and anticyclones of similar growth rate. Synoptic experience suggests however, and theory can offer an explanation, that the cyclonic system develops the more rapidly, that the surface convergence is not fully balanced by divergence within the wave train but is spread laterally over a wide area: thus the baroclinic zone tends to take on a dominantly cyclonic character, especially so over the oceans where the moisture supply provides a source of energy in ascending motion.

Typically, we observe old occluded cyclones stagnating in a cold trough north of the baroclinic zone and becoming merged with succeeding travelling lows, the ridge between being squeezed out of existence without producing a high centre. It remains then to determine the circumstances in which the anticyclonic development pattern shall have time to assert itself and these appear to depend very much upon the suitable spacing, and phases of development, of neighbouring depressions. It is a truism to say that the cyclone series must be broken or diverted but it is meaningful to ask whether the primary agency can be assigned to the active cyclonic cells, to the anticyclonic development itself, or to a wider control in the upper long-wave pattern. We hope to return to this question later but meanwhile we may observe that vigorous anticyclonic building across the baroclinic zones seems usually to be <u>preceded</u> by a stagnating depression further west although the self-developing and self-maintaining properties of the anticyclone may later become dominant.

These are however problems of anticyclonic development in general and are not dependent on whether a new high centre happens to appear in the process.

Table I

New anticyclonic centres - 1950 - 51

Total Number of Cases - 42

		Dates	Location	Remarks
I	Simple upper oscillation in baroclinic zone with pre-existing anticyclone to the south Number of cases - 4	20.10.50	E. Atlantic	Illustrated in Figure.
		4.11.50	N. of Br. Isles	But initially lower pressure to South : almost Type II.
		11. 8.51	C. Atlantic	Tendency towards a double frontal system
		26.10.51	Scandinavia	Soon absorbed the high to South.
II	Simple upper oscillation in baroclinic zone but without pre- existing high to south of anti-cyclogenetic region Number of cases - 21	10. 6.50	Br. Isles	Pre-existing cold depression Portugal : See Figure
		2. 7.50	Br. Isles	Low Spain to N. Africa
		8. 7.50	France	Low NW Africa Azores high
		12. 7.50	France	Low NW Africa Azores high
		18. 7.50	France	Low NW Africa Azores high
		25. 7.50	France to Germany	Low NW Africa. Weak Oscillation
		19. 8.50	France to Germany	Low NW Africa. Azores high Weak Oscillation.
		29. 8.50	N. Spain to France	Low NW Africa. Azores high Weak Oscillation.
		21.11.50	S. of Greenland	Cut off cold low to South
		26.11.50	Br. Isles to Germany	Cut off cold low Azores
		1.12.50	S. of Greenland	Cut off cold low Azores Large oscillation. See Figure.
		28. 2.51	France	Low Med. Large oscillation.
		16. 4.51	E. Atlantic	Cut off cold low Azores
		19. 4.51	North Sea	Cut off cold low Biscay.
		29. 5.51	N. of Br. Isles	Cut off cold low Biscay.
		17. 8.51	SW of Br. Isles	Low NW Africa
		14. 9.51	N. Spain to France	Low NW Africa
		16. 9.51	SE of Greenland	Cut off cold low NW of Azores
		4.11.51	Scandinavia	Large cold low Br. Isles
		15.11.51	N. Scandinavia	Large cold low E. Atlantic
		9.12.51	E. Atlantic	Low Azores
III	Anticyclonic disruption of upper wave pattern Number of cases - 9	8. 3.50	E. Atlantic	Cold low formed to W of Portugal
		26. 3.50	E. Atlantic	Cold low moved to W of Portugal. See Figure
		20. 4.50	Baltic	Cold low cut-off Mediterranean
		12. 6.50	SW of Iceland	Cold low cut-off N. of Azores
		30. 8.50	Iceland	Cold low cut-off Br. Isles
		12. 2.51	N. of Br. Isles	Cold low cut-off Biscay
		23. 3.51	C. Atlantic	Cold low cut-off S. of Azores
		25. 4.51	C. Atlantic to SW Iceland	Cold low cut-off Azores
		4. 9.51	France to Germany	Cold low cut-off Italy

/IV

Table I (cont'd.)

	Dates	Location	Remarks
IV Complex	20. 2.50	C. Atlantic	Double structure. See Figure
Number of cases - 8	29. 7.50	Jan Mayen	Low formed E. of Greenland
	4.11.50	N. of Br. Isles	Double structure, lows to N and S
	26.12.50	S. Scandinavia	
	20. 3.51	Br. Isles	Almost Type II
	20. 6.51	C. Atlantic	Double structure, almost Type III
	11. 7.51	S. of Iceland	Small scale feature
	28.10.51	C. Atlantic	Small scale. Double structure

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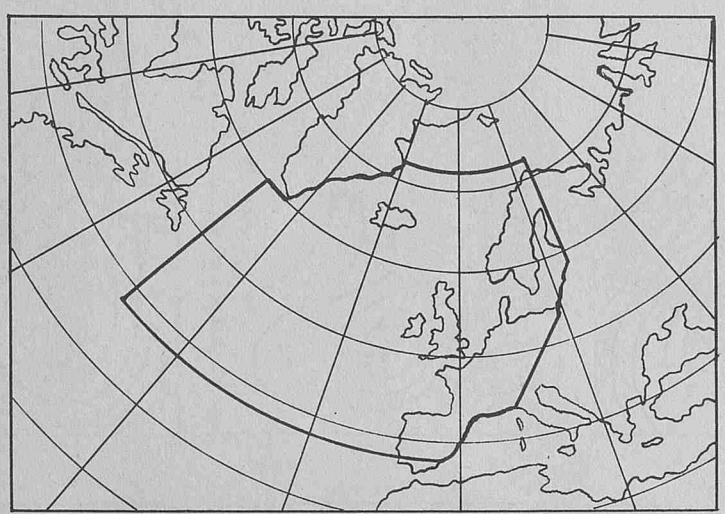


Figure 1. Area studied for new anticyclonic cells.

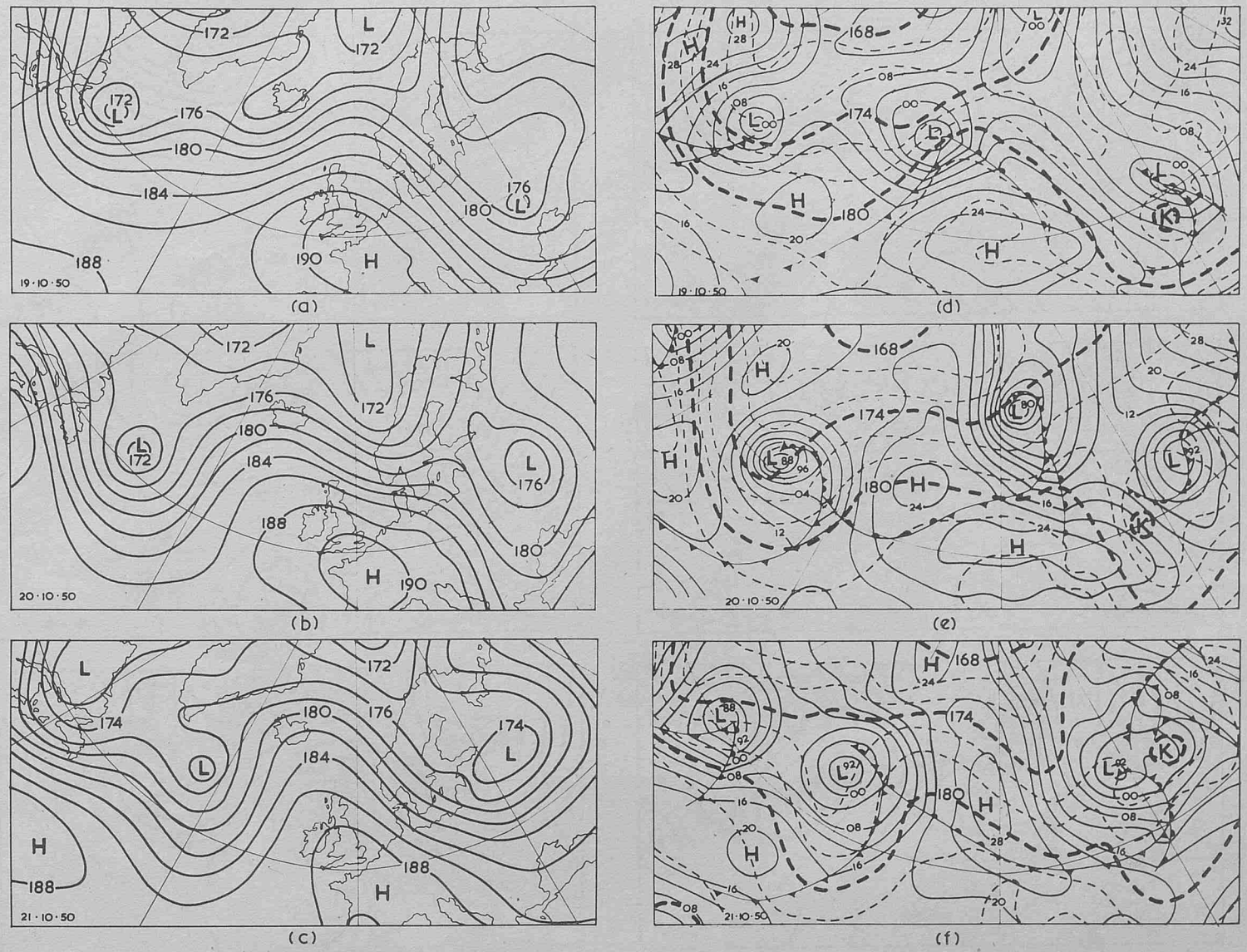


Figure 2. Anticyclonic development with simple oscillation in the upper westerlies 19 - 21 October 1950. for discussion see section 3·1

(a), (b), (c) – 500mb. Contours 0300 G.M.T.

(d), (e), (f) – Surface isobars and fronts 0000 G.M.T. full lines. 1000 - 500mb. thickness 0300 G.M.T. broken lines.

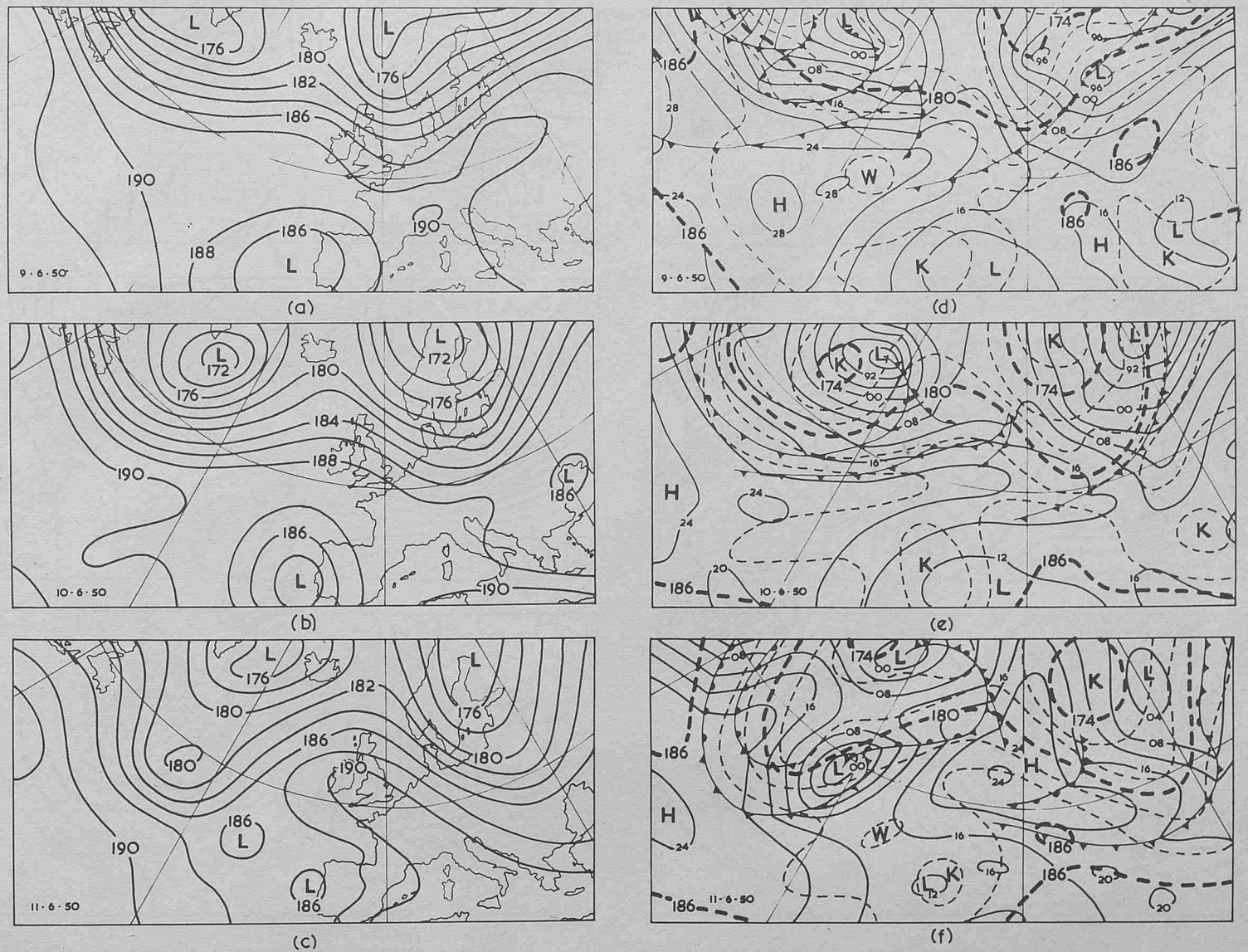


Figure 3. New anticyclone with oscillation in upper westerlies north of pre-existing depression. 9–11 June 1950. for discussion see section 3·2

(a), (b), (c) — 500mb. Contours 0300 G.M.T.

(d), (e), (f) Surface isobars and fronts 0000 G.M.T. full lines. 1000–500mb. thickness 0300 G.M.T. broken lines.

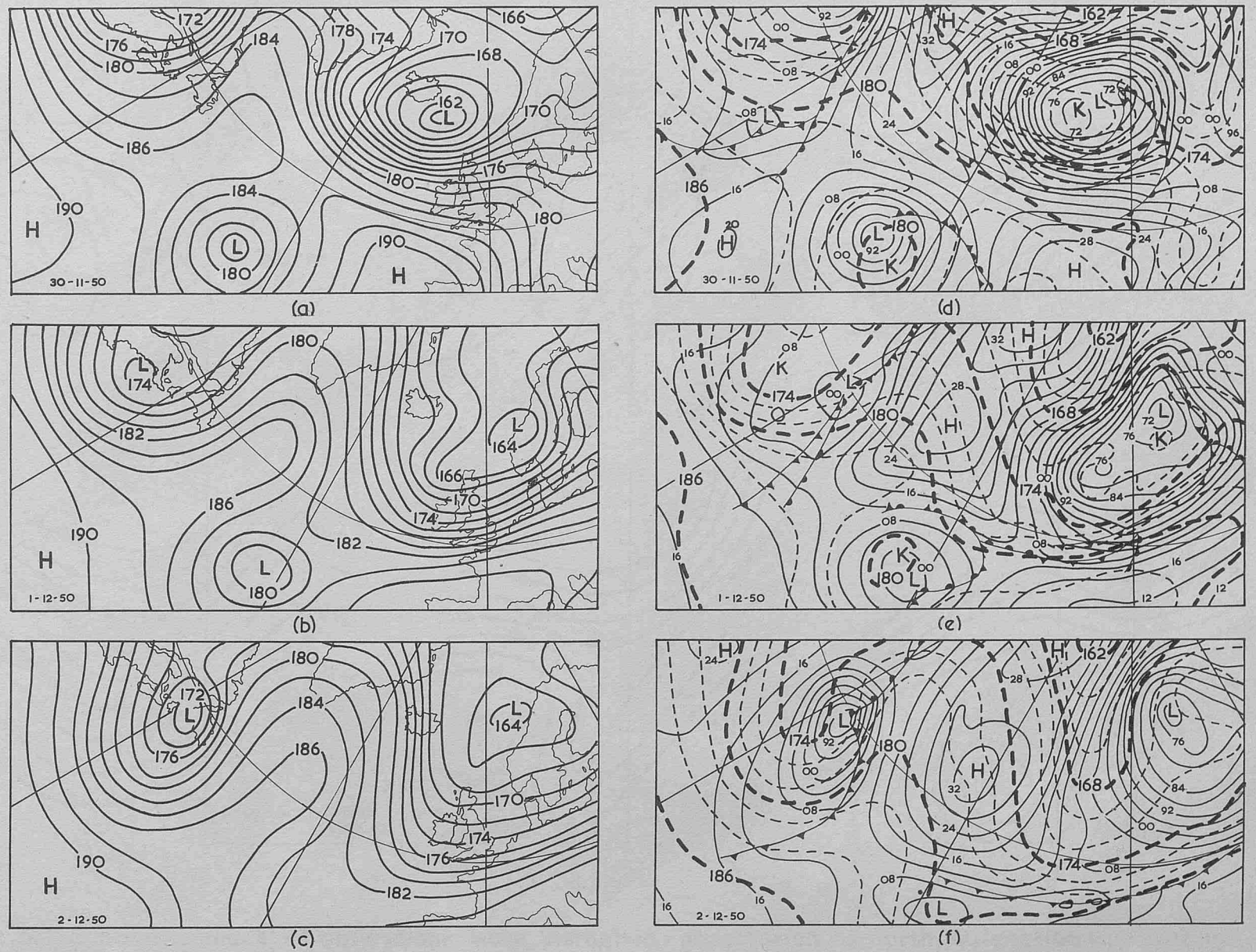


Figure 4. New anticyclone with oscillation in upper westerlies north of pre-existing depression.
30 November — 2 December 1950. for discussion see section 3·2.

(a), (b), (c) — 500mb. Contours. 0300 G.M.T.

(d), (e), (f) Surface isobars and fronts 0000 G.M.T. full lines.
1000–500mb. thickness 0300 G.M.T. broken lines.

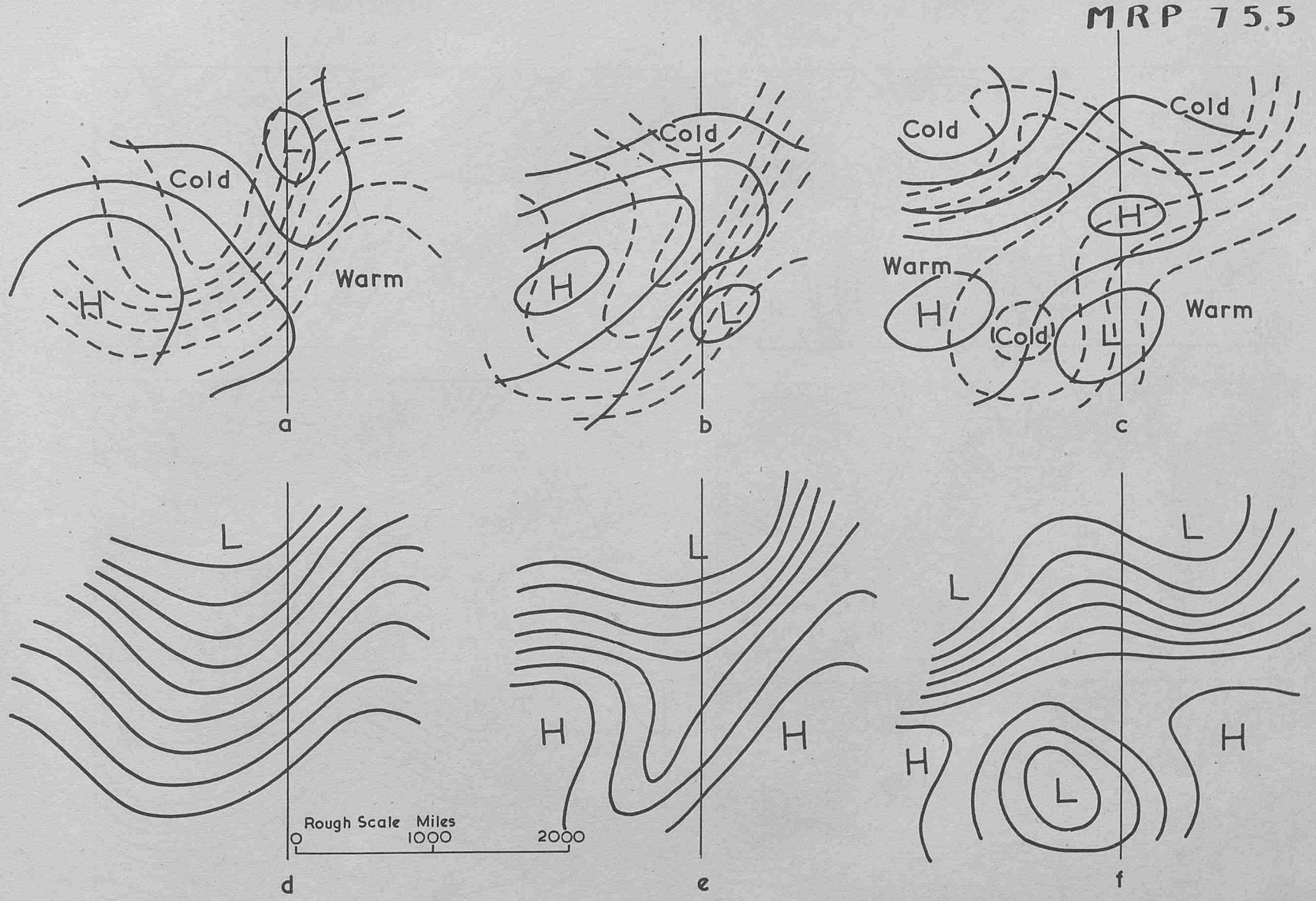
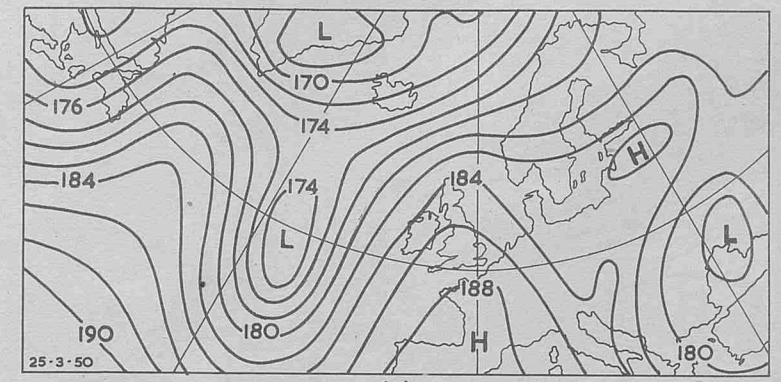


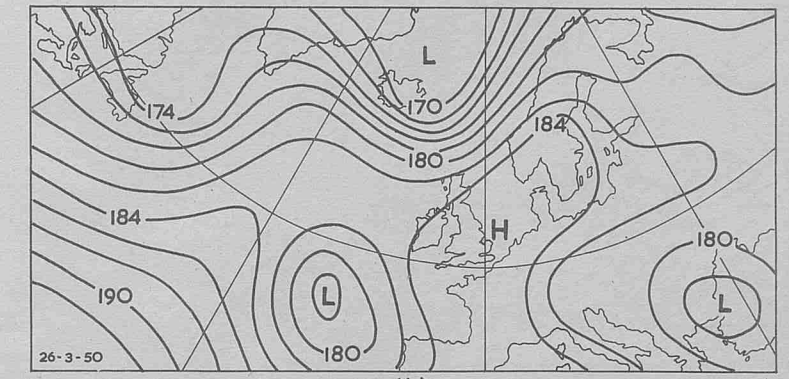
Figure 5. Schematic model of anticyclonic wave disruption: progression to north, stagnation and cutting-off to south with surface ridge building across the neck.

(a), (b), (c), Surface isobars full line – 8 mb. intervals; 1000–500 mb. thickness – 200 ft. intervals.

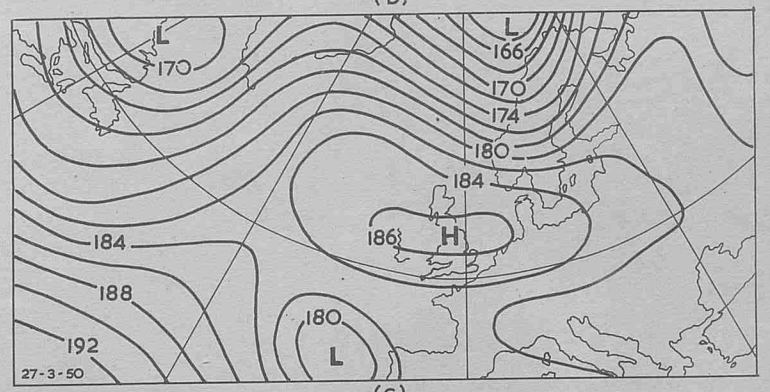
(d), (e), (f). 500 mb. contours – 200 ft. intervals.



(a)



(b)



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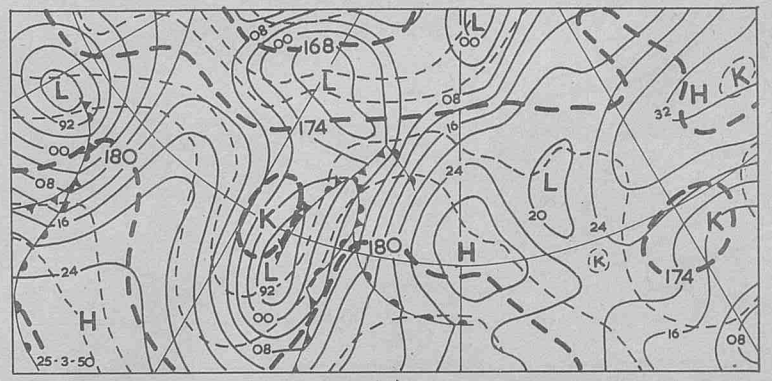

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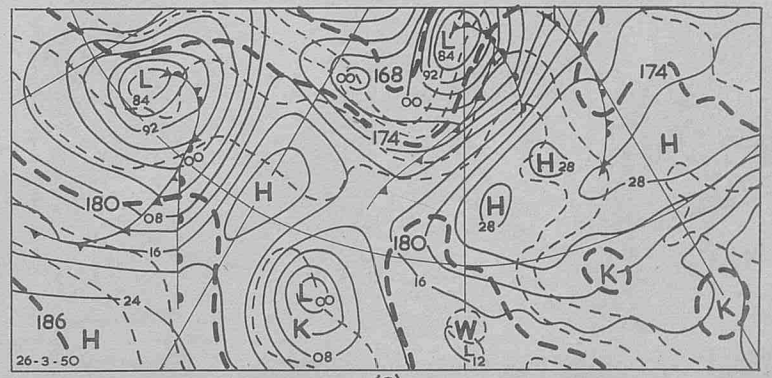
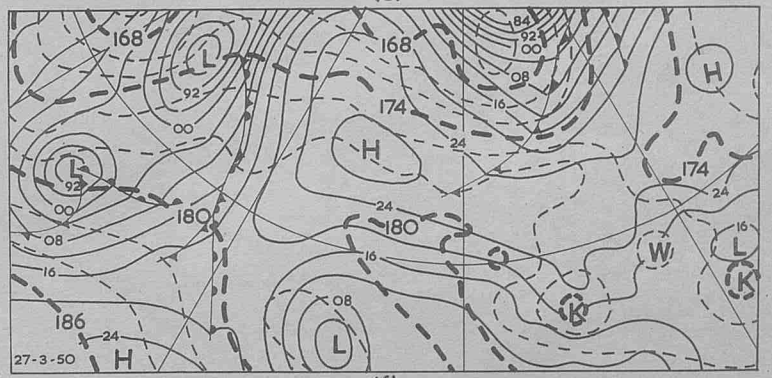


(e)



(f)

Figure 6. Anticyclonic Disruption 25 - 27 March 1950. for discussion see section 4.

(a), (b), (c) — 500mb. Contours 0300 G.M.T.

(d), (e), (f) Surface isobars and fronts 0000 G.M.T. full lines.
1000-500mb. thickness 0300 G.M.T. broken lines.

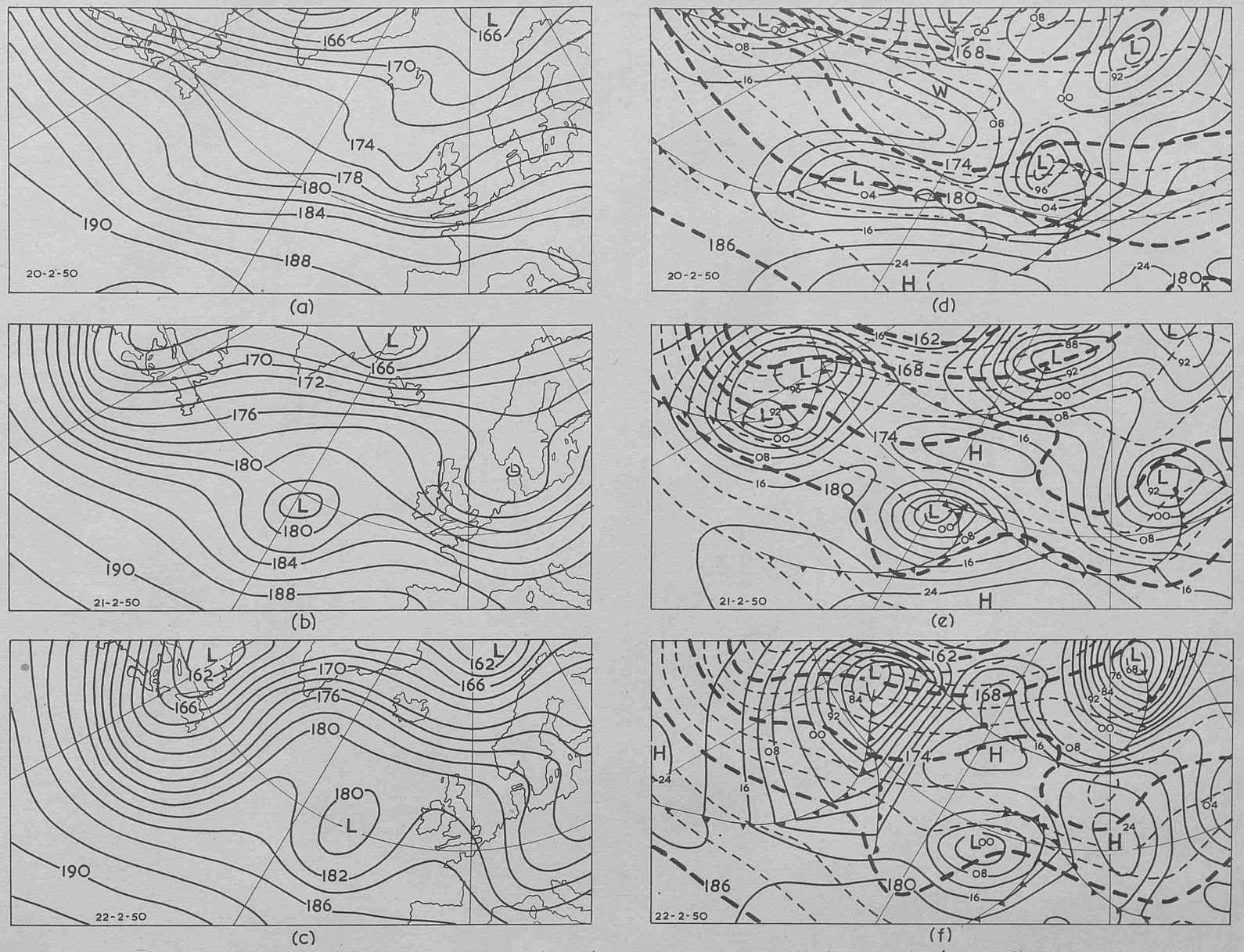


Figure 7. New anticyclone in complex situation with two series of baroclinic (frontal) depressions. 20 – 22 February 1950. for discussion see section 5.

(a), (b), (c) — 500 mb. Contours. 0300 G.M.T.

(d), (e), (f) Surface isobars and fronts. 0000 G.M.T. full lines: 1000 – 500mb. thickness 0300 G.M.T. broken lines.

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M.R.P. 757
S.C.III/141
23rd September, 1952.

MAY 27 1953

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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Meteorological Report on the Singapore Cloud Detection Trials

by

R. Frost (Meteorological Office)

1. Introduction

During the period 18th April to 18th June, 1950, the Royal Air Force Transport Command Development Unit carried out a test of certain airborne radar equipment for cloud detection using for this purpose a Lancaster aircraft based at Singapore.

The special equipment used in this investigation and described in detail in (1) included (a) a three cm. radar for cloud detection, (b) a recording accelerometer mounted as close to the centre of gravity of the aircraft as possible for recording vertical accelerations and (c) an auto-observer containing an altimeter, air speed indicator, clock and two microammeters, the latter being used in conjunction with pairs of electrostatic probes projecting about 8 inches outside the aircraft and being insulated from it, one pair being mounted at the wing tips and the other at the nose and tail of the aircraft, for measuring the abeam and fore and aft components of the external electric field.

To avoid waste of flying time in searching for areas of cumulonimbus cloud, daily flight plans were drawn up in consultation with the station meteorologists at the Royal Air Force Station, Changi, Singapore, and flights were generally made in the afternoon when cumulonimbus activity was most pronounced.

In order to carry out the cloud investigation, runs were generally started at a distance of about 50 miles from the cloud so that the maximum range of the radar detection could be ascertained, and normally the run culminated in a penetration of the cloud through the region giving the maximum radar response. Occasionally after emerging from a cloud a rapid 180° turn would be made and a second penetration of the cloud carried out without the preliminary run in from 50 miles. On one occasion 7 penetrations were made into one cumulonimbus cloud during a period of approximately 80 minutes.

During the run in, meteorological observations and photographs of the cloud were taken by the navigator, corresponding records of the radar display being made by the radar operator. The accelerometer was switched on from about two minutes before the entry of the aircraft into the cloud until after the exit of the aircraft from the cloud. Times of entering and leaving the cloud were recorded by the navigator and similar observations in respect of the radar response area were made by the radar operator.

Qualitative estimates of the type and intensity of the precipitation and of the degree of turbulence encountered during the penetration were made by the captain and recorded by the radar operator.

The flying technique used during the penetrations was similar to that developed by the U.S.A.A.F. during the Thunderstorm Project (2) and consisted in keeping the aircraft straight and level by reference to the gyro compass and artificial horizon, using the minimum of control, and flying at a reduced turbulence speed usually about 135 kt. indicated air speed, no account being taken of violent changes of altitude and indicated airspeed unless the safety of the aircraft was involved.

During this investigation 87 penetrations were made into the cores of 49 different cumulonimbus clouds, 19 penetrations were made into 18 cumulus clouds and 4 penetrations were made into unidentified clouds.

The observations of meteorological interest made during the course of this investigation and which are the first of their kind in the Tropics are summarised and discussed in this report. Comparison is also made with observations in the U.S.A. (2) and with observations in England by Jones (3) in co-operation with aircraft from the Royal Aircraft Establishment at Farnborough (3).

2. Heights of cumulus and cumulonimbus clouds

On the run in towards the cloud the angle of elevation of the top of the cloud was measured with a bubble sextant, the range of the cloud being simultaneously observed on the aircraft radar. A number of such observations was normally made over about 10 miles of the run in and the mean of such observations taken, so that with a knowledge of the aircraft's altitude the height of the cloud could be determined. The scatter of the calculated heights from the mean indicated that on about 95% of all occasions the mean would be in error by less than 1,100 feet.

The mean heights of 23 clouds classified in accordance with the International Cloud Atlas and measured in this way are exhibited in ascending order of magnitude in Table I. It was not possible to measure the tops of all clouds which were subsequently penetrated as frequently they were obscured.

Table I. Measured cloud tops over Sumatra and Malaya

Date	5/5	27/5	10/6	10/6	27/5	*26/5	**10/6	19/5	30/5	31/5	1/6
Local Time	1721	1511	1540	1557	1527	1641	1400	1523	1445	1800	1402
Type	c	c	c	c	c	cb	c	cb	cb	cb	cb
Mean height in feet	12920	21300	21780	27180	28706	29200	32830	34000	36970	37370	38310
No. of observations	1	2	4	2	1	1	5	2	1	6	5

Date	31/5	2/6	30/5	2/6	1/6	4/5	10/6	4/5	1/6	1/6	1/6	2/6
Local Time	1515	1435	1510	1508	1750	1238	1330	1655	1525	1615	1715	1640
Type	cb	cb	cb	cb	cb	cb	cb	cb	cb	cb	cb	cb
Mean height in feet	38570	40236	40600	41660	41770	43310	44100	44500	47960	48730	48760	54600
No. of observations	6	6	3	5	5	1	5	1	7	7	7	4

* Almost certainly a dissipating cumulonimbus cloud cf. Section 7(a).
** Probably in process of transformation into cumulonimbus cf. Section 3.

It can be seen from Table I that the change over from cumulus to cumulonimbus cloud occurs at about 30,000/33,000 feet, and that many of the cumulonimbus clouds have tops well over 40,000 feet, the highest occurring at 54,600 feet.

These agree very closely with Van Bemmelen's calculations in 1912 (4) from photographs of cumulonimbus cloud over Batavia which showed that the tops of cumulonimbus cloud were in many cases over 33,000 feet and on occasions reached to 50,000/53,000 feet.

The significance of these limits is a matter of some interest. As cumulonimbus cloud is distinguished from cumulus cloud by the observed fact that the tops of the former are beginning to acquire a fibrous appearance, this implies that there is a rapid increase in the number of ice crystals in the Tropics at 30,000/33,000 feet where the temperatures are about -30/-35°C. with a standard deviation of 2.3°C. This is in very good agreement with laboratory experiments. Thus Findeisen and Schulz (5), using an expansion chamber with a volume of air of two cubic metres in order to obtain adiabatic rates of cooling comparable to those prevailing in convective clouds, found that with an expansion equivalent to a vertical velocity of 5 m.sec.$^{-1}$ there was a very rapid increase in the numbers of ice crystals at -35°C. whilst with a more rapid expansion equivalent to a vertical velocity of 20 m.sec.$^{-1}$, the sudden increase in the numbers of ice crystals occurred at a temperature of -29°C.

The upper limit of the cloud tops, at 55,000 feet approximately, is in very good agreement with the observed height of the tropopause in the Tropics indicating that, as might be expected, the tropopause constitutes an effective upper barrier to vertical convection in the Tropics.

3. Radar response from clouds

The intensity of the radar response or echo from a volume of water drops has been shown by Ryde (6) to be proportional to

$$\Sigma N D^6$$

where N is the number of drops of diameter D per unit volume, so that one raindrop of diameter 2 mm. can give an echo equivalent to that from a million cloud particles of diameter 0.2 mm. Thus the echo from a precipitating cloud, or from a cloud containing drops of raindrop size suspended in the vertical currents, is far greater than that from a non-precipitating cloud, and in general echoes are only detectable from clouds which contain drops of raindrop size. It follows therefore that the radar picture of a cloud only indicates that portion of a cloud which contains drops of raindrop size, and that normally this does not coincide with the visual appearance of the cloud. From the times of entry into and exit from both the clouds and the radar response areas within the clouds in the Singapore flights the mean diameter of the visual cumulonimbus cloud was found to be approximately 6 miles, whilst that of the responding portion of the cloud was approximately $4\frac{1}{2}$ miles.

Radar echoes from cumulonimbus clouds were generally very sharp, compact and bright. They could be detected at ranges of about 40 miles, the maximum range being 47 miles, and as the range decreased the cumulonimbus clouds sometimes showed the existence of two or more response areas.

Radar echoes from cumulus clouds were more diffuse and irregular on the whole, although this was not an invariable rule and depended upon the size and age of the cloud. Towering cumulus cloud with a vertical development of 25,000/30,000 ft. could be picked up at about 30 miles range, the maximum being 35 miles, whilst smaller cumulus clouds could sometimes be detected at 5/10 miles.

No echoes were received from clouds other than cumulus or cumulonimbus clouds.

At very short ranges it was found that the echoes from cumulonimbus clouds generally retained their discrete and solid quality whilst those from cumulus clouds in general became very wispy and diffuse. This was not always the case however and it was noted in (1) that occasionally echoes from cumulonimbus clouds became more nebulous at close range, whilst at other times cumulus echoes became more solid. It is very probable that in the former the clouds were dissipating and that in the latter the cumulus clouds were in the process of development into cumulonimbus clouds.

/It

It is of interest to note that the cumulus cloud which is reported as becoming more solid at close range is that given in Table I with a top of 32,830 ft. and which is almost certainly developing into a cumulonimbus cloud. Similarly one cumulonimbus cloud with a top of 36,970 feet which is quoted as becoming wispy is almost certainly dissipating, as on the subsequent penetration it was found that the turbulence was light, that the precipitation was slight and that there was a slight down-draught throughout the complete traverse.

It was noted in (1) that simultaneously with the entry of the aircraft into the clear cut edge response of a cumulonimbus cloud the pattering of rain or hail could be heard on the radome even above the engine noise, which indicated that in cumulonimbus clouds the precipitation had quite sharp and vertical edges, whereas in cumulus clouds giving fuzzy responses there was a horizontal gradient of precipitation intensity.

From the few occasions on which observations on hail and snow were available it appeared that the radar echoes from hail were indistinguishable from those from heavy rain, whilst those from snow appeared to be less bright and solid than those from rain.

There is one significant difference between the American and Singapore observations, for whereas in the former echoes were only obtained from cumulus clouds whose tops extended from 2,000/5,000 feet above the freezing level, in the latter echoes were obtained from cumulus clouds whose tops were well below the freezing level. In one developing cumulus cloud whose top was 12,920 feet (or 2,000/3,000 feet below the freezing level) which was penetrated at 5,000 feet, the aircraft flew in the response area for nearly two miles and reported the precipitation as light rain.

This difference in the radar response of cumulus clouds in mid latitudes and in the Tropics is consistent with observed differences in rainfall from cumulus clouds in mid latitudes and the Tropics.

In mid latitudes the observational evidence all seems to support the Bergeron (7) theory of precipitation and the proponents of this theory categorically state that all moderate to heavy rain is released from clouds which extend above the freezing level.

In tropical latitudes however it has now been established that moderate to heavy rain can be released from clouds which are well below the freezing level. In the Caribbean, for example, moderate to heavy rain has been observed to fall from cumulus cloud whose tops are at 9,000/10,000 feet (8) whilst Langmuir (9) quotes reports from American pilots which show that heavy showers can occur over tropical islands in the Pacific from clouds which are well below the freezing level (apparently even from cumulus clouds whose tops were as low as 5,000/6,000 feet). To account for this Langmuir suggests that precipitation in the Tropics is caused by a chain reaction process which depends upon the up-draught velocity in the cloud, the thickness and water content of the cloud and the diameters of the droplets within the cloud.

4. Hail

Hail which, although relatively infrequent, is generally regarded as one of the worst hazards of flying in cumulonimbus cloud gave surprisingly little trouble during the Singapore flights. In this series of flights hail was encountered on 7 out of a total of 87 penetrations into cumulonimbus cloud and these are listed in Table II.

On May 1st hail was encountered in the same cumulonimbus cloud in two penetrations separated by an interval of 16 minutes.

/Table II.

Table II. Penetrations encountering hail

Date	Local Time	Height of Cloud in feet	Mean Up-Draught Rate (ft./sec.)	Height of Aircraft in feet	Temperature outside cloud corrected for air speed	True air speed (kt.)	Duration of Precipitation (sec.)	Notes on Precipitation	Other remarks if any
1.5.50	1512	-	0	11,500	8°C	172	180	Moderate hail, rain	Same cumulonimbus cloud, lightning.
1.5.50	1528	-	+8	12,500	7°C	175	60	Moderate to heavy hail	
4.5.50	1618	-	0	15,000	-4°C	180	30	Light hail	No lightning.
31.5.50	1704	-	+2	11,900	10°C	167	135	Light to moderate rain, light hail	Up-and down-draughts in cloud. Lightning
1.6.50	1620	48,730	+3	13,600	6°C	172	222	Moderate snow then hail	Up-and down-draughts causing vertical displacements of ±500 ft. Cloud dissipated 20 min. later, no lightning.
2.6.50	1551	-	-2	10,200	10°C	162	80	Light hail	Up-and down-draughts. Lightning.
3.6.50	1554	-	0	11,200	7°C	170	20	Light rain and hail	Penetration 6 min. earlier gave large up-and down-draughts including one of 53 ft./sec. Lightning.

One other occasion of hail was noted in a cloud of somewhat dubious type which was described as "dark cu. and overcast", details of which are included in Table IV.

/In

In the Florida thunderstorms discussed in (2) hail occurred on 3 out of 131 penetrations at 11,000 feet and on 13 out of 128 occasions at 16,000 feet, whilst in the Ohio thunderstorms hail occurred on 26 out of 217 occasions at 10,000 feet and on 10 out of 208 occasions at 15,000 feet.

The reporting of both up-draughts and down-draughts in the cumulonimbus clouds despite the fact that the mean up-draught and down-draught currents over the whole traverses were negligible and the fact that in the only cloud in which it was possible to determine the height of the cloud top, the cloud top was 48,730 feet, indicate that the hail occurred during the later stages of cumulonimbus development prior to the dissipating stage.

On only one occasion in the Singapore investigations was it possible from an examination of the records of individual clouds to evaluate the up-draught over a limited portion of the cloud, and this gave a mean up-draught speed of 53 feet per second over a distance of about 4,000 feet.

Now it is usually supposed that strong upcurrents are required in order to support hailstones and the magnitudes of the up-draughts required may be estimated from the calculated terminal velocities of hailstones of specified magnitude. Thus for hailstones of 1cm. and 2cm. diameter respectively, the terminal velocities are 40 feet/sec. and 53 feet/sec. respectively.

In the American investigations (2), studies of the rates of vertical growth of the radar echoes from four cells in which hail was observed showed that in every case the up-draught speed was in excess of 50 feet per second, whereas in 3 cells in which it was certain that hail did not occur the rates of growth were less than 40 feet per second. It is possible that the infrequency of hail encountered during the Singapore flights, and in the Tropics generally, is due to the chain reaction precipitation mechanism which, by removing liquid water from the rising air prior to the transformation of a cumulus cloud into a cumulonimbus cloud, hinders the subsequent growth of the ice crystals falling from above.

Thus according to Mason (10) if ice crystals fall into a dense supercooled cloud containing many large droplets, the collisions between the ice crystals and the droplets (which then become frozen) are so frequent that the ice crystals quickly become irregular roughly spherical aggregates which develop into hailstones if the upcurrents of air support them in the cloud long enough, whereas if the ice crystals fall through supercooled clouds with low liquid water content, collisions are infrequent and the growth of the ice crystals proceeds predominantly by diffusion so that richly branched crystals develop and ultimately snowflakes are formed.

5. Snow and Sleet

Snow and/or sleet was encountered on 14 out of 110 penetrations and these are listed in Table III.

/Table III.

Table III. Occasions of Snow and Sleet.

Date	Local time	Type	Height of Cloud Top	Mean Up-draught rate in ft./sec.	Height of aircraft in feet	Corrected air temp. outside cloud	True air-speed in kt.	Duration of precipitation in sec.	Notes on precipitation	Other remarks if any
4.5.50	1628	Large dark cu. and overcast?	-	0	13000	1°C	174	40	Hail & sleet	
5.5.50	1509	Unidentified	-	0	17000	-2°C	189	75	Snow & sleet	
5.5.50	1520	St	-	0	17000	-2°C	183	15	Sleet	
17.5.50	1316	Cb	-	+ 9	11000	+9°C	170	72	Sleet	
19.5.50	1548	Cb	35000	+ 2	16000	+1°C	178	40	Light rain and snow	
19.5.50	1612	Cb	-	- 2	15000	0°	168	120	Moderate then heavy rain some snow	
23.5.50	1702	Cb	-	0	12500	6°	162	30	Moderate rain then sleet	Cloud reported dissolving rapidly on further penetration 6 min. later with mean down-draught of 30 ft./sec.
31.5.50	1523	Cb	38570	- 5	10000	10°	162	75	Light to moderate rain and snow	Cloud reported dissolving with down-draught on penetration 22 min. later
31.5.50	1824	Cb	37370	- 4	10600	8°	163	130	Light rain and sleet	Two subsequent penetrations both gave light to moderate rain and down-draughts
1.6.50	1532	Cb	48000	+ 4	14400	4°	168	82	Light rain moderate snow	Same cloud
1.6.50	1551	Cb	48000	+ 2	14000	6°	166	100	Snow	Multi-cored cb
1.6.50	1628	Cb	48000	- 7	12500	6°	168	95	Snow	Just below base of cb. cloud dissipating with squall line
1.6.50	1620	Cb	48730	+ 3	13600	6°	172	222	Moderate snow then hail	
1.6.50	1644	Cb	48730	- 1	12000	8°	167	85	Snow and heavy rain	Cloud reported dissipating Same cloud

/For

For comparison, in the Ohio thunderstorms there were 55 occasions of snow and sleet out of a total of 208 penetrations at 15,000 feet and 4 out of 217 at 10,000 feet.

In the Singapore flights it was not practicable on account of engine heating to make a habit of flying above the 0°C. level which was about 15,000 feet, but penetrations above this level were made on a few occasions, the two highest being at 17,000 feet. In all cases of penetrations made at or above the 0°C. level, light snow or sleet was encountered, but on no occasion was any ice picked up.

This at first sight appears somewhat surprising, for as the concentration of liquid water at any height in a convective cloud is determined mainly by the temperature of the cloud base and the vertical distance above the cloud base, it had been expected on the basis of computations made by Best (11) that the highest rates of ice accretion would occur in the Tropics.

The reasons for this lack of icing may be either (a) that the time spent by the aircraft in the core of the cumulonimbus cloud, where high liquid water concentrations should be found, was insufficient for ice to be picked up, or (b) that severe icing in the Tropics is only found in towering cumulus clouds just before they reach the cumulonimbus stage, for it is known that the formation of ice crystals in cumulonimbus clouds leads to a rapid reduction in the liquid water concentration and Lewis (12) states that the transformation from liquid water drops to snow particles proceeds very rapidly once it begins, frequently reducing the amount of the former by about 90% in 20 minutes. It is also probable that the risk of severe icing in the Tropics is reduced by the chain reaction precipitation mechanism.

In the Ohio thunderstorms no icing was experienced on 72% of occasions at 15,000 feet and in the Florida thunderstorms no icing was experienced on 71% of occasions at 16,000 feet. It is probable however that on many occasions in the latter flights the 0°C. level was higher than 16,000 feet.

6. Lightning

Lightning was experienced on 35 out of 110 penetrations, but no strikes on the aircraft occurred. During one penetration, according to (1), eighteen flashes were counted which were sufficiently close for the heat radiated to be felt by the crew.

From the American investigations the main conclusions on lightning were that

(a) lightning was never found in small cumulus, but that as the cumulus grew into the cumulonimbus stage lightning began.

(b) the cloud tops reached a height where the environment temperature was about -21°C before the first lightning occurred.

(c) as the intensity of the storm increased the frequency of lightning increased.

(d) the maximum cloud height and maximum lightning frequency occurred together.

(e) as the height of the thunderstorm decreased the frequency of the lightning decreased.

From the Singapore observations the number of occasions on which lightning was reported in cloud when details of the cloud top height were also available are given in Table IV.

/Table IV.

Table IV. Frequency of lightning

Height range in thousands of feet	5/15	15/25	25/35	35/45	45/55
No. of clouds within height range	1	2	5	11	4
No. of occasions of lightning	Nil	Nil	Nil	4	2

Table IV indicates that the lightning risk is nil with cloud tops below 35,000 feet and that the chance of lightning is about one in two or three when the cloud tops are between 35,000 and 55,000 feet whilst according to the pilot, the lightning flashes appeared to be more frequent in cumulonimbus clouds which were passing through the dissipating stage.

On the not unreasonable assumption that the cumulus top is rising at about 500 feet per minute, cf. Section 7, comparison of Tables I and IV indicates that the first lightning occurs about 5/10 minutes after the cloud had passed into the cumulonimbus stage which supports the view of Simpson and Scrase (13) that the existence of large numbers of ice crystals is a prerequisite for the separation of electric charges.

The microammeters connected to the electrostatic probes were observed to oscillate wildly inside the clouds, giving readings up to $\pm$200 microamperes. In general no readings were obtained unless the aircraft was actually inside the cloud, and the presence or absence of lightning had no observable effect.

7. Turbulence within cloud

As in (2), for convenience, turbulence in clouds in the Singapore flights is discussed under two sub-heads (a) draughts and (b) gusts.

In a draught the height of the aircraft varies because of the mean upward and downward motion of the air and hence, provided an aircraft is flown at constant power setting, attitude and airspeed, draught velocities can be measured by the rates of vertical displacement of the aircraft as recorded by the altimeter. Gusts are local irregular variations of velocity which subject the aircraft to a series of sharp accelerations without systematic change of altitude and these can be measured by an accelerometer.

The distinction between draughts and gusts however is not always clear cut, for the edges of some up-draughts may be sufficiently sharp to produce an acceleration, whilst some gusts may cause changes in the attitude of an aircraft and hence produce changes in the altitude. Furthermore the manner in which a pilot exercises control can effect both changes in altitude and accelerations.

As various types of aircraft do not react in the same way to different gusts a gust formula is required in order that the accelerometer records from different aircraft may be compared and aero-dynamicists have given a formula for the instantaneous vertical velocity which, superimposed upon a steady horizontal wind, would produce an acceleration of a given magnitude upon a particular aircraft which is flown on straight and level flight.

This 'sharp edged' gust formula is

$$\text{Equivalent gust velocity} = \frac{2wn}{\rho_0 aKV_1}$$

where w is the wing loading, n is the acceleration increment, ρ_0 is the density of the air at ground level, V_1 is the indicated air speed, a is the slope of the lift coefficient and K , which is called the alleviation factor, is an empirical factor which varies with the wing loading.

It can be seen from the above formula that for the same aircraft and the same equivalent gust velocity, doubling the velocity of the aircraft doubles the magnitude of the acceleration increment, so that in order to reduce the stress on the aircraft and the apparent bumpiness it is desirable to fly through cumulonimbus clouds at as low a speed as possible, and this procedure was adopted in all three investigations.

The highest acceleration increment experienced in the Singapore investigation by a Lancaster aircraft flying at 140 kt. was 1.1g which corresponds to an equivalent gust velocity of 49 feet per second. In the English investigation (3) the highest acceleration increment experienced with a Spitfire aircraft flying at 180 kt. was 1.5g which corresponds to an equivalent gust velocity of 51 feet per second, whilst in the American investigation (2) with a P.61 flying at 180 m.p.h. the highest acceleration experienced was 1.8g which corresponds to an equivalent gust velocity of 73 feet per second. It is of interest to note that this last acceleration increment occurred on the edge of a strong up-draught in which the pilot temporarily lost control of the aircraft.

(a) Draughts

Continuous altimeter records throughout the cloud unfortunately were not made in the Singapore flights, so that in general only mean up-draught and down-draught speeds computed from the altimeter readings on entry to and exit from the cloud and from the time spent within the cloud could be determined, and these are given in Table V. In this table up-draughts are counted as positive.

Table V. Distribution of mean up-draughts and down-draughts

Up-draught speed in feet/sec.	21	18	15	12	9	6	3	0	-3	-6	-9	-12	-15	-18	-21
Cumulus cloud	-	-	-	1	2	4	3	5	2	-	-	-	-	-	-
Cumulonimbus	1	1	1	2	4	6	18	28	16	4	-	-	-	-	-

With the reservation that the cumulonimbus data are five times as numerous as the cumulus data, Table V shows that with few exceptions the characteristic feature of a cumulus cloud is, as might be expected, the preponderance of up-draughts although the maximum up-draughts were obtained in cumulonimbus clouds.

As the dissipating stages of cumulonimbus clouds are known to be marked by down-draughts, cf. for example Table III, it would appear that the maximum up-draughts are reached soon after the transition from cumulus to cumulonimbus.

The auto-observer was used in all penetrations after the 32nd and readings of the altimeter, air speed and microammeters were recorded photographically, exposures being taken before and during the traverses at rates varying from 1 to 4 per minute. An examination of these records showed that, despite the precautions taken, considerable variations in the air speed occurred at times indicating that the aircraft was not always maintaining constant attitude, so that except possibly qualitatively little use can be made of these records for determining mean up-draught speed over limited portions of the cloud.

It is of interest to note that on the occasion in which the aircraft made seven penetrations into a single cumulonimbus cloud in a period of 80 minutes, an examination of the altimeter records on entry to and exit from the cloud supplemented by the auto-observer records showed that during the first few penetrations over a period of 35 minutes there was a steady up-draught. On the 5th penetration 15 minutes later slight up- and down-draughts were encountered in the first half of the traverse followed by an up-draught in the second half. On the 6th penetration 12 minutes later there was a down-draught followed by a slightly stronger up-draught, whilst on the 7th penetration 17 minutes later there was a steady down-draught throughout the cloud traverse.

/In

In one hailstorm, to which reference has already been made, the navigator recorded that the aircraft entered the cloud at a height of 11,100 feet, was carried in a down-draught to 9,900 feet and then in the space of 15 seconds was lifted to 10,700 feet which indicated that there was a mean up-draught speed of 53 feet per second over a horizontal distance of about 4,000 feet.

On one day two penetrations separated by an interval of 17 minutes were made into a developing cumulus cloud. During the first run in, the height of the cumulus top as measured at 1511 was 21,300 feet, and the aircraft entered the cloud at 15.15.35 at 13,000 feet and emerged from the cloud at 15.16.25 at 13,400 feet. During the second run in, the height of the cloud as measured at 1527 was 28,706 feet and the aircraft entered the cloud at 15.32.50 at 12,500 feet and emerged from the cloud at 15.34.45 at 13,500 feet. It was unfortunate that on the second run in only one measurement of the cloud top could be made as the navigator reported that the top of the cumulus appeared to be blowing away. In both penetrations precipitation was encountered and reported as light to moderate rain.

It can be seen that the mean up-draught speed on the first traverse was 8 feet per second and on the second traverse 8.7 feet per second whilst the vertical velocity of the cloud top was 7.7 feet per second.

This lack of increase of the vertical velocity is not in accordance with the model of a developing cumulus cloud in mid latitudes as suggested in (2) and if correct indicates that in the Tropics the temperature in a cumulus cloud does not differ greatly from the temperature of the environment.

On 26th May, 1950 two penetrations separated by an interval of 10 minutes were made into a cumulonimbus cloud whose range of detection by radar was only 8 miles instead of 30/40 miles cf. Section 3.

During the first penetration which was made at about 9,000 feet slight up- and down-draughts were experienced with a meanup-draught over the whole traverse of 1 foot per second approximately whilst during the second penetration at approximately the same level there was a consistent down-draught, the mean over the traverse being 3 feet per second.

On the first run in the height of the cumulonimbus top as measured at 1642 was 29,200 feet, and on the second run in at 1650 was 27,895 feet. Unfortunately only single measurements of the cloud top were made on both run ins so that the computed rate of descent of approximately 2 feet per second is subject to considerable error.

The general agreement between the observations however indicates that the cumulonimbus cloud was dissipating.

(b) Gusts

The conclusions of the American investigations (2) and Jones' investigation (3) conflict on the gust aspect of turbulence within cumulonimbus clouds. According to (2) "one of the characteristics of a cell (as shown by the radar response) is the turbulence therein, consequently a minimum of turbulence or a region of no turbulence occurs at the boundary of each cell, and usually turbulence ceases entirely for a distance of 2,500 feet when an aircraft passes through the boundary of a cell," whereas according to (3) "about half the number of flights through cumulonimbus clouds in England showed consistent turbulence throughout the cloud, whilst the large majority of the flights in the other half gave a noticeably rougher patch on the approach to or inside the response area. Only in about 25% of all the flights was the roughest patch encountered wholly within the response area."

An analysis of the Singapore accelerometer records was carried out for both the responding and non-responding portions of the cloud in order to investigate this disagreement.

Table VI shows the frequency of acceleration increments $\geq |0.2|g$ experienced by the Lancaster aircraft. In this table all acceleration increments have been included irrespective of the height of the aircraft, which normally flew between

10,000 feet and 12,000 feet, but on occasions flew between 8,000 feet and 17,000 feet, as over this height range the variation of gustiness appeared to be slight, and no attempt was made to correct for the small variations of air speed by means of the equivalent gust formula.

Table VI. No. of acceleration increments in cloud $\geqslant 0.2g$

Acceleration increment	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0	1.1
Responding area of cloud	1111	472	238	65	22	10	6	1	5	1
Non-responding area of cloud	473	211	72	22	8	4	4	2	-	-

As the time spent in the response area of the cloud to that in the non-response area of the cloud was in the ratio 3 : 1 (cf. Section 3) Table VI indicates that there is little difference in the degree of turbulence in the responding and non-responding portions of the cloud.

This is also borne out by an inspection of Figure I which shows the total number of accelerations $\geqslant 0.2g$ experienced in the responding and non-responding portions of the clouds on each flight.

On one occasion the aircraft flew through a cumulonimbus cloud entirely between two responses, and on three other occasions flew in and out of the responses in multicored cumulonimbus.

The number of acceleration increments $\geqslant 0.2g$ experienced whilst flying in the non-responding portions of such clouds are shown in Figure I, whilst typical accelerometer records in which the turbulence is greatest in the response area, non-response area and in the regions between responses respectively are shown in Figures 2(a), 2(b) and 2(c).

These figures and Table VI lend little support to the American conclusions and appear to be in general agreement with Jones' findings.

It will be noted that Figure 2(d) which shows the portion of an accelerometer record when the aircraft was flying through a down-draught followed by an up-draught of 53 feet per second, the latter extending for a distance of about 4,000 feet, does not appear to support the American suggestion that thunderstorm gusts are whirling pieces of air driven by and superimposed upon the thunderstorm draughts, or their conclusion that such gusts are strongly correlated with the draughts.

It may be that the American theory of gusts is applicable to the relatively simple build-up stage of non-frontal convective cloud in mid latitudes, for in this case, as the response area denotes the region of maximum water content and as it is in the region of the up-draught that large quantities of water vapour are converted into liquid water, it might be expected that there should be some correlation between the radar response and the up-draught, although it would seem that as suggested by Jones (3) the eddies should occur on the boundaries of the up-draught area and not within it. The lack of correlation between the turbulence and the radar response in the Singapore investigation may be due to the fact that the majority of flights were made into cumulus or cumulonimbus clouds in which down-draughts due to density differences caused by rain had already begun and suggests that if any gusts were originated by the initial up-draught such gusts had persisted long after the original up-draught causing them had ceased.

Figure I is instructive, for bearing in mind that in these flights the pilot in general selected and penetrated the most impressive looking cumulonimbus cloud which he could find, deliberately flying through the portion of the cloud giving

the greatest radar response, it indicates that it is impossible for a pilot to estimate from either the visual appearance or the radar picture of a cumulonimbus cloud the degree of turbulence which he is likely to experience. This agrees with the conclusions of (1) in which it was stated "During the trials attempts were made to associate the turbulence encountered on penetration with some previous characteristic of the radar response. These attempts failed and no obvious method of forecasting turbulence could be observed either by range of initial detection, size brilliance or any other quality of the response or by the vertical extent of the responding portion of the cloud".

Since, however, little or no turbulence was encountered outside a cumulus or cumulonimbus cloud the radar technique suggested in (1) for avoiding turbulent areas was to avoid all radar responses by about 1 mile thereby ensuring that the cloud itself was also avoided. Apart from the question of turbulence this procedure also means that areas of heavy precipitation would be avoided and the risk of icing reduced.

Jones (3) noted that significantly greater numbers of positive accelerations occurred as the aircraft entered the response area whilst the reverse occurred as the aircraft emerged from the response area and the present records were investigated to find whether this also held for convective cloud in the Tropics.

All positive gusts $\geqslant 0.5g$ were selected and their times of occurrence after the aircraft had entered the response area were noted. Table VII shows the number of such accelerations at various times after entry into the response area.

Table VII. No. of positive accelerations $\geqslant 0.5g$ within response area

Time in secs. after entry into response	0-4	5-9	10-14	15-19	20-24	25-29	30-34	35-39	40-44	45-49	50-54
No. of accelerations	13	11	9	6	6	6	5	6	3	4	4

Table VIII shows the number of negative accelerations $\geqslant 0.5g$ at various times before emergence from the response area.

Table VIII. No. of negative accelerations $\geqslant 0.5g$ within response area

Time in secs. before exit from response	0-4	5-9	10-14	15-19	20-24	25-29	30-34	35-39	40-64
No. of accelerations	0	3	2	2	2	1	1	1	0

Table VII is consistent with Jones' findings, but Table VIII is inconclusive. This latter, however, may be due to the difficulty of determining the time of exit from the response.

It is uncertain whether this feature of the accelerations is a characteristic of conditions within the response area of a cumulonimbus cloud or whether it is connected with the manner in which the pilot exercises control.

The crew's impression of the degree of turbulence during the flight was noted as slight, light, moderate or heavy, but it was found that these impressions were highly subjective, and turbulence which was considered moderate by one individual was often considered light by another, whilst turbulence which was considered moderate in early penetrations was frequently considered as light or even slight on later penetrations when more experience had been gained.

A comparison of the accelerometer records with the pilot's qualitative estimates of turbulence suggests that it is the frequency of acceleration increments above a certain amplitude rather than the maximum amplitude which determines the sensation of turbulence and in the Lancaster 5 acceleration increments of 0.5g in 10 seconds was considered heavy turbulence.

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8. Temperatures within clouds

The temperatures in cloud as registered by an ordinary direct reading thermometer and uncorrected for the speed of the aircraft were in all cases equal to or lower than the uncorrected temperatures outside the cloud, and in general the fall of temperature on entering the cloud was correlated with the intensity of the precipitation experienced within the cloud.

The number of occasions of temperature falls of specified amounts on entering the cloud are shown in Table IX.

Table IX. Temperature falls on entering cloud.

Environment temperature minus cloud temperature	0°C	1°C	2°C	3°C	4°C	5°C	6°C
No. of occasions	6	5	17	16	10	9	1

Somewhat similar results were obtained by Barret and Riehl (14) who gave an explanation suggested by Langmuir that outside the cloud the thermometers were warmed dry adiabatically due to the partial stagnation of the air at the measuring element, whereas inside the cloud there was a mixture of liquid water and air, of which the former is nearly incompressible, so that if the same dynamic heating were applied to the temperatures outside and inside the cloud the latter would apparently be low.

If the full dynamic heating correction is applied to the temperatures outside the cloud and none is made to the temperatures within the cloud then on the average the difference of temperature between the cloud and the environment is very small which does not appear unreasonable, cf. Section 7.

On the other hand however in two penetrations which were made into developing cumulus cloud, the temperatures as measured within the cloud were 5°C lower than the uncorrected temperature of the environment or, which appears somewhat paradoxical, the temperatures within a developing cumulus cloud were approximately 1°- 2°C lower than the true temperature of the environment.

It is possible that on these occasions and on other similar occasions of large temperature differences the thermometers had become completely wetted so that they were effectively functioning as wet bulb thermometers, and some support to this suggestion is afforded by the fact that on 5 occasions on which the cloud was penetrated at or near the 0°C level and snow was encountered the uncorrected temperature differences were smaller and varied between 0°C and 2°C.

9. The life cycle of a tropical convective cloud

From (2) it appeared that a thunderstorm or large shower cloud consisted of one or more cells, each of which was a unit of convective circulation which had a life cycle of 1 to 3 hours more or less independently of adjacent cells. In this life cycle there were three stages, the cumulus or building stage, the cumulonimbus mature stage and the cumulonimbus dissipating stage.

Owing to the diminished importance of the Bergeron precipitation mechanism in the Tropics, the feature which the Americans noted as the identifying feature of the transition between the first two stages, namely the occurrence of rain at the surface, is not appropriate in the Tropics.

Instead it is suggested that in the Tropics there should be a slight modification and that the three stages should be (a) the cumulus or building stage (b) the active cumulonimbus stage which can be recognised by the fact that the protuberances of the cumulus cloud have softened and have acquired a fibrous appearance and (c) the dissipating cumulonimbus stage which is recognisable by the formation of the anvil.

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In stage (a) a general up-draught prevails and the horizontal gradient of precipitation intensity suggests lateral mixing with the environmental air. The time taken to go through this stage appears to be about $\frac{1}{2}$ to 1 hour. In stage (b), which is reached when the cloud tops have reached a height of about 30,000 to 33,000 feet, the mixing with the environment appears to be only of secondary importance. During this stage, which also lasts for about $\frac{1}{2}$ to 1 hour, the cloud reaches its maximum height which may extend to the tropopause, up-draught and down-draught speeds and turbulence reach their maximum and hail and lightning may occur.

Stage (c) develops from (b) when the down-draughts become predominant and ultimately there is a gentle down-draught throughout the lower levels of the cloud with only slight turbulence. This stage does not appear to last for long, in general for less than $\frac{1}{2}$ hour.

Conclusions

The Singapore flights in addition to adding to and clarifying our knowledge of convective clouds in the Tropics, in general confirm the conclusions of the Thunderstorm Project although in some important respects they conflict.

In particular they show that

(a) contrary to experience in middle latitudes radar echoes in the Tropics are received from cumulus clouds whose tops are well below the freezing level indicating the presence of drops of raindrop size in such clouds and suggesting that some precipitation mechanism other than that implied by the Bergeron theory is required;

(b) the transition from cumulus to cumulonimbus cloud occurs in the Tropics at a height of 30,000/33,000 feet where the temperatures are $-30^{\circ}C/-35^{\circ}C$ which appears to support laboratory experiments of the rapid increase of ice crystals in an expansion chamber at such temperatures;

(c) lightning appears unlikely to be encountered with cloud tops below 35,000 feet;

and suggest that

(d) the risk of hail and severe icing in cumulonimbus clouds in the Tropics may be slight.

The Singapore trials lend no support to the American conclusions on the relationship between turbulence and radar response and in general support the East Hill findings.

The conclusions on turbulence from these trials is that practically all turbulent areas can be avoided if the radar response areas are avoided by a distance of about 1 mile.

Acknowledgement is due to the Research Branch, Transport Command, and to the pilot and crew of the T.C.D.U. who at considerable risk and discomfort made the very careful series of meteorological observations on which this report is based.

/References

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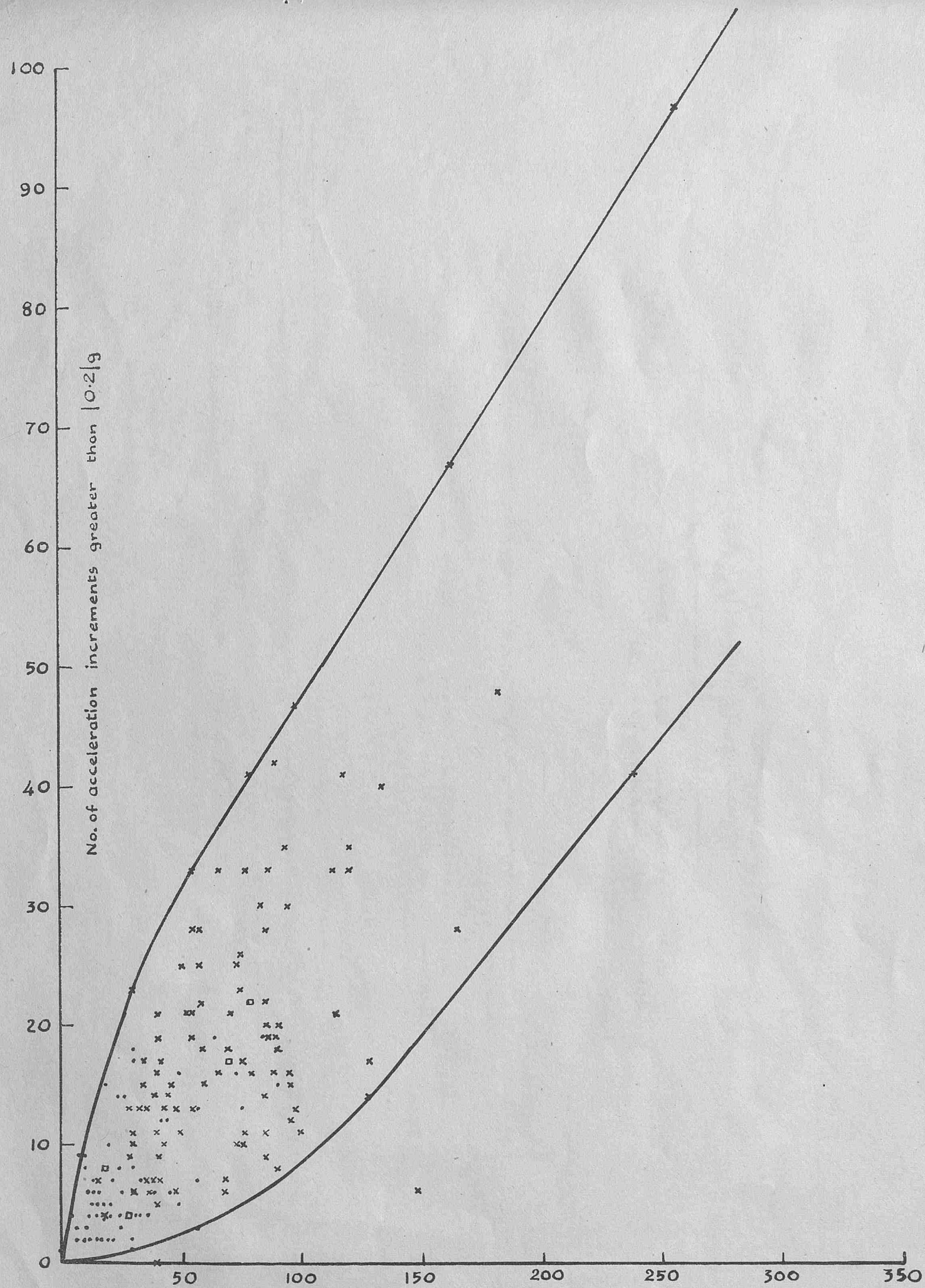

Fig I The number of acceleration increments $\geq |0{\cdot}2|g$ as a function of time within the responding or non responding portions of a cloud.

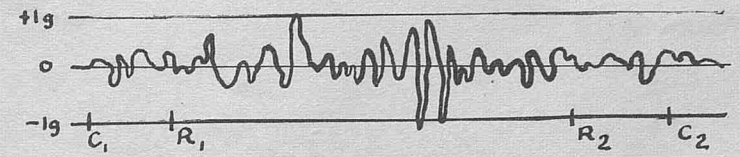

Fig. 2(a). Accelerometer Record 16·5·50. In Cloud 16h.40min. 5sec. Scale
Out of Cloud 16h. 41 min. 25sec. ⟵⟶=10sec.

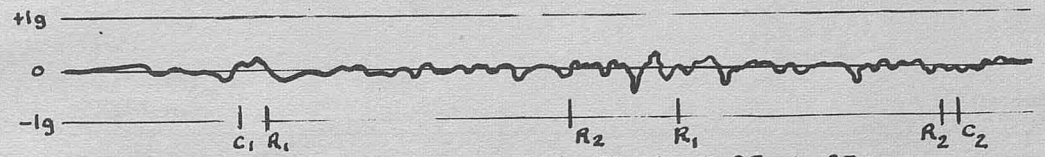

Fig. 2(b). Accelerometer Record 30·5·50 In Cloud 15h.23min. 25sec. Scale
Out of Cloud 15h. 25 min. 00sec. ⟵⟶=10sec.

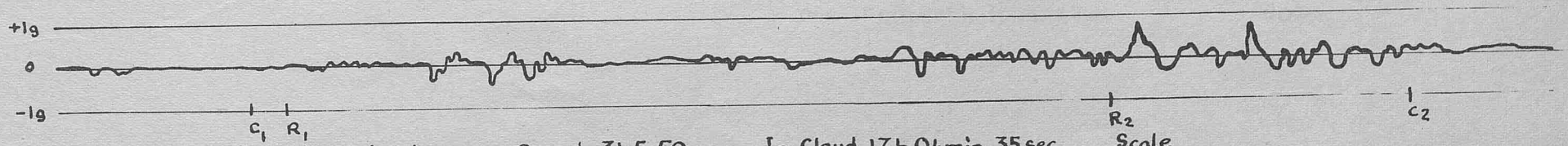

Fig. 2(c) Accelerometer Record 31·5·50 In Cloud 17h.04min. 35sec. Scale
Out of Cloud 17h.07min. 45sec. ⟵⟶=10sec.

C_1 Cloud begins R_1 Response begins
C_2 Cloud ends R_2 Response ends

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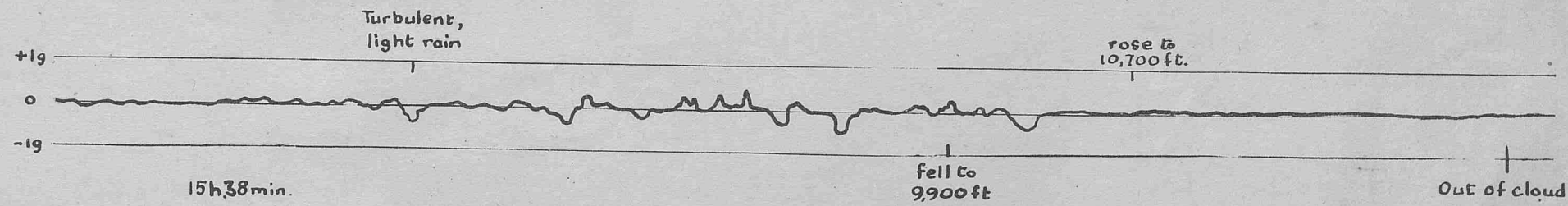

Fig. 2(d). Accelerometer Record 3·6·52

In Cloud 15h. 35min. 35sec.

Out of Cloud 15h. 39min 45sec.

Scale ——— =10sec.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No ..... by Mr. ..... A copy is available in the Library of the Meteorological Office."

M.R.P. 758
S.C.II/122
26th September, 1952.

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The Storm of 14-15th January, 1952 as a moving vortex

by
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The intense storm of 14-15th January, 1952[1] presented a good example of vortex structure. The purpose of this note is to put certain data on record, with comments on the surface structure (paras. 2, 4, 5, 6) and on the upper air conditions and their relation to depression theory generally (paras. 3, 7, 8).

2. Fig. 1 shows the distribution of surface wind speed in relation to distance from the isobaric centre. In deriving the means of wind speed for this diagram, a system of weighting was used (in cases where two or more wind observations were available at a given distance from the centre) to give double weight at each station to the supposed best exposed direction. Means were then taken for the three stations (Lerwick, Stornoway, Tiree) for given distances of the station from the low pressure centre. For comparison the Vr product 14525 (kt. x km.) is shown on the same diagram. It will be seen that from about 240 km. to 800 km. from the centre the mean wind speed closely follows the vortical distribution. On the other hand, within 240 km. of the centre, the mean values available indicate a close approximation to "solid rotation".

3. Fig. 2 is intended to show the wind structure of the depression both in relation to distance from centre and in relation to height. It repeats the diagram of surface wind speed and shows, for a larger area to southward, the upper winds at the 825 and 375 mb. levels, taken as the means of the 800 and 850 mb. levels and of the 350 and 400 mb. levels respectively. In the case of the 375 mb. level, on account of the great change of wind speed with time (referred to in para. 7 below), the observations have been divided into two time groups, namely 14d. 21h. and 15d. 3h. in diagram A, and 15d. 9h. and 15h. in diagram B. In diagram B, unfortunately, lack of upper air observations makes it impossible to continue the curve nearer to the centre than about 600 km. Both of the curves for the 375 mb. level are subject to a good deal of uncertainty, owing to the insufficiency of observations and the rapidity of the changes in progress at these high levels. The curve for the 825 mb. level is a mean through a considerable number of observations, showing however a good deal of scatter towards the outer (right hand) end. The left hand extension of curve A for 375 mb., considered in relation to the surface curve, brings out the great shear of wind (referred to later at para. 7) in the vertical over the central area of the depression during the period of rapid deepening. Owing to the absence of upper wind data for Lerwick, all three curves outside a radius of about 600 km. relate necessarily to positions lying well to southward of the centre of the depression and in a relatively restricted sector.

4. In Geoph. Mem. No. 79[2] an analysis is given of the (surface) vortical features of 20 depressions. In the majority of these, the wind distribution outside a certain radius (which averages 420 km.) from the isobaric centre approximated to Vr = Const. Inside the distance just mentioned, the wind

/speed

speed distribution was roughly $V \propto r$. In an exceptional case, namely the depression of October 26-28, 1936, the vortex was developed up to some 280 km. from the centre, giving a maximum wind speed (hourly mean) at this distance of 57 knots against an average of 42 for the twenty depressions. The suggestion was made that abnormally high wind speeds are in general associated in this way with continuation of the vortex up to an unusually close approach to the centre.

The storm of 14-15th January, 1952 surpassed the one just referred to in that the vortex is developed up to 240 km. from the centre, giving at this point the quite abnormal (surface) mean speed of 60 kt.

The Atlantic storm of 26-27th December 1951 (of "Flying Enterprise" fame) was even more intense. It is discussed by P.R. Brown in "Marine Observer", July 1952[3], on the basis of observations made by the Ocean Weather Ship "Weather Observer". He shows that the vortex was developed up to about 225km. from the isobaric centre, giving at this point a surface speed of 70 kt. with gusts of at least 100 kt. (Brown calls attention to the fact that the higher speed is partly accounted for by the ratio of surface wind to gradient being higher at sea than even at coastal stations such as are used in the present paper).

5. Interesting vorticity points arising are -

(a) The mean vortical intensity in the outer region of the 20 depressions referred to in para.4 was 18,340 (kt. x km.), that of the exceptional depression of October 26-28th, 1936 was 16,380, that of the still more outstanding case of January 14-15th, 1952 only 14,525 and that of the depression of 26-27th December, 1951 probably about 15,000.

(b) On the other hand the rates of spin of the central area vary in the opposite sense, namely one revolution in about 35 hours, once in 17 hours, once in 14 hours and once in 11 hours respectively. (Vorticities 0.37, 0.76, 0.90 and 1.12hr^{-1} relative to earth's surface.)

(c) The spin of once in 14 hours is incidentally almost exactly equal to $2\omega \sin\varphi$ in lat. 60°N, so that the central air mass of the storm is going round relatively to the earth's surface at twice the rate that the earth's surface is itself going round in space.

(d) It will be noted that when the rate of spin in a depression centre falls to $2\omega \sin\varphi$ - or about twice the value usual in a moderately deep depression - the cyclostrophic and the geostrophic terms (V^2/r & $2\omega V \sin\varphi$) in the gradient wind equation becomes equal. With further fall the geostrophic predominates.

6. The inference from 5(a) and (b) above would seem to support the view suggested in G.M.79, i.e. in a young depression, as over the Atlantic, the central area is small and the rate of spin high, whilst the vortical intensity of the outer vortex is low. With age - i.e. in general as the depression comes over the British Isles - the radius of the central area increases, the rate of its spin diminishes and the vortical intensity in the outer region rises. The extreme in the matter of spin is perhaps the tropical hurricane of the West Indies with a revolution of once in 4 or 5 hours, a central area radius of 50-90 km. and an outer vortical intensity of perhaps only 7,000. It may be that, more often than we yet think, one cyclone lives through the whole series of types from tropical hurricane to decaying depression of high latitudes. It is tempting to think of the principal cyclone systems of the Earth's atmosphere as having features analogous to the disturbances on the sun, but with the annual term

/predominant

predominant in the periodicity of the tracks. In the Atlantic, the tracks of the basic disturbances would reach their maximum each year in the West Indies about September, passing gradually to higher latitudes, possibly the latitude of the British Isles in October to late November when the lowest pressures often occur in Britain, to still higher latitudes in late December to late January when extreme wind speeds often occur in Britain. But obviously there is great complexity and the discussion of these possibilities digresses too far from the subject of the present paper. (A calculation for the storm of January 14-15th, 1952 on the basis of Shaw's revolving fluid theory places the"kinematic centre" 73 km. to the left of the path of the isobaric centre and the "tornado centre" 59 km. to the right of the path. These distances are smaller than are usually found. The mean values for the 20 depressions referred to earlier are 163 and 90 km.).

7. It will be useful to look at the kinetic energy in the surface layer. If the depression is of the simple vortex type with maximum velocity Vm at distance R from centre, it can be shown that the total kinetic energy per unit thickness of the surface layer[4] is

$$\pi \rho \, R^2 V_m^2 \left(\frac{1}{4} + \log_e \frac{S}{R}\right)$$

where S is the outer radius to which the depression affects the atmosphere. In the depressions examined in Geoph. Mem. 79, S was of the order of 2R. In this case the second term - which represents the energy outside the radius R- is 2.76 times the first term which represents the energy of the inner part in solid rotation. The following table gives values of $R^2V^2m/10^6$ a convenient term proportional to the total kinetic energy -

	$R^2V^2m/10^6$	
Mean of 20 depressions		337
Seven Depressions of Apr.-Sept.	186)	
Thirteen " " Oct.-Mar.	441)	
Seven " " Dec.-Feb.	582)	
Depression of 26-28 Oct. 1936		269
" " 14-15 Jany.1952		211
" " 26-27 Dec. 1951		225
W.Indian Cyclone 80 kt. at 90 km.		52

The separation of the depressions of the British Isles into seasons shows that these develop their greatest surface level kinetic energy in mid-winter. The last case shows that the larger and more violent type of W.Indian Cyclone has about one fourth of the total kinetic energy in surface levels of a young and vigorous Atlantic depression, or one seventh that of the average large depression which has come over the British Isles and is approaching the stage of occlusion. It would seem that the older the depression the wider the area over which the surface kinetic energy is distributed.

Assume that the energy of the surface circulation - (say) the lowest 3 km. - is taken from the energy of a jet stream in the upper 3 km. of the troposphere, over a width of the stream equal to 2R i.e. the diameter of the central area of the depressions, and in the time of 12 hours. Calculation shows that a reduction in the mean speed of the jet stream by about 10 kts. would suffice. Even if the width of jet stream from which the kinetic energy is taken is reduced by half, and if some allowance is made for other losses, a reduction of perhaps 30 kts. in mean speed of a stream with a mean of 120 and perhaps a maximum of 160 kts. would still suffice.

There were in fact high winds (a jet stream) over the central area of the depression of January 14-15th 1952 at the time when the depression was deepening rapidly. The rapid deepening took place between 14d. 12h. and 15d. 6h. At 14d. 3h. there is 116 kt. at 400 mb. at Weather Ship Station Item and 113 kt. at 200 mb. at Jig. At 14d. 15h. there is 151 kt. (266°) at Item at 300 mb. Again at Item, on the north side of the centre, the speed ranges from 84 to 108 knots (236°) from 400 mb. to 250 mb. (the highest point reached in the ascent) at 14d. 21 h. whilst below the 700 mb.

/level

level it nowhere exceeds 11 kt. (260°). At the same time at Jig, on the south side of the centre, the speed is 38 kt. (270°) at the surface but 144 kt. (237°) at 300 mb. Thus there was also very high shear of wind in the vertical direction whilst the depression was deepening.

On the other hand, so far as can be seen, the winds at the 300 mb. and the higher levels, later on, namely at 15d. 9h., probably did not exceed about 90 kt. over the central area of the depression and the shear between the 300 mb. and 700 mb. levels had greatly diminished. These are the sort of changes which might be expected if the depression in lower levels is dissipating part of the energy of the jet stream.

8. From other examples it would seem that high wind speed and high shear over the central area may not in themselves be a sufficient prior condition to the formation of or the deepening of a depression, though frequently - perhaps invariably - they are present when such a development takes place. From the circumstance that the deep depressions of winter in temperate latitudes originate or deepen generally over the sea it would seem probable that some convection from the surface layers is also essential. It should give the necessary interchange between surface layers and the high speed currents of the upper troposphere. It is also true that when high wind speed and high shear in the upper troposphere are absent, as is often the case in summer in temperate latitudes, large travelling depressions do not form, however great the convection. The evidence therefore is consistent with a dual cause for depressions, the first and principal condition being high speed in the upper troposphere, i.e. a jet stream, and high shear; the second condition - possibly not much more than trigger action - being convection from the surface levels.

In the case of the depression of 14-15th January, 1952, both of the conditions necessary for deepening or development appear to have been satisfied. The vortical structure in the surface layers is further consistent with development produced by the evacuation of air from a central area.

To say that a jet stream is a necessary condition for the formation or deepening of a travelling depression is the same thing as saying that the tropopause must have a rather sharp step-down in level in the transition from warm to cold air. In a developing, fast travelling depression with wide-open warm sector, the tropopause step should be situated some distance to north of the depression centre, whilst the jet stream should be almost over the centre. When occlusion takes place, the tropopause step (or part of it) and correspondingly the jet stream (or a branch of it) should be transferred southward, so that the principal necessary condition for the formation of another depression (i.e. a secondary) tends to be reproduced further south in some measure whenever a parent depression is occluded; the right hand extension of curve B of Fig. 2 is in keeping with this. But the second necessary condition for a secondary (convection) might be wanting, e.g. if the occlusion edge lay over the land in winter.

Looked at in this way, it is the planetary circulation which is the primary cause and source of energy of depressions; the depression is the first stage of turbulent mixing which continually re-endows the surface layers of air with momentum which they gradually lose again, partly through ground friction and partly through conversion of kinetic to potential energy in the occlusion process; and every depression tends to leave, somewhere in upper levels, the preliminary condition for the birth of yet another depression. Adverting to the remarks of para.6, there may be, in any given season and year, some feature of the tropopause structure and upper troposphere circulation which locates the basic track and which tends to persist.

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Fig.1 - Depression of 14-15 January 1952

Surface Winds

(Weighted means - Lerwick, Stornoway, Tiree.)

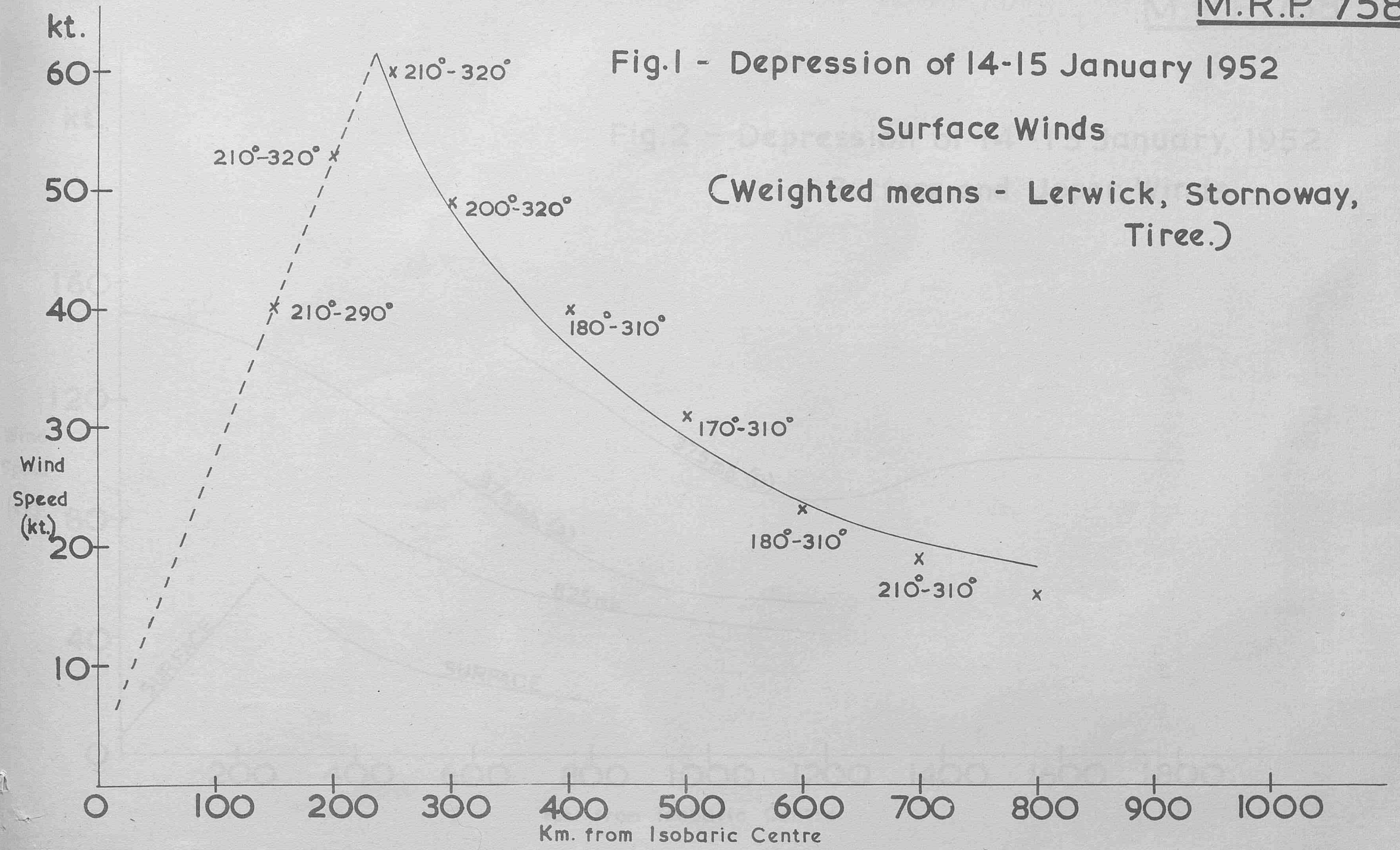
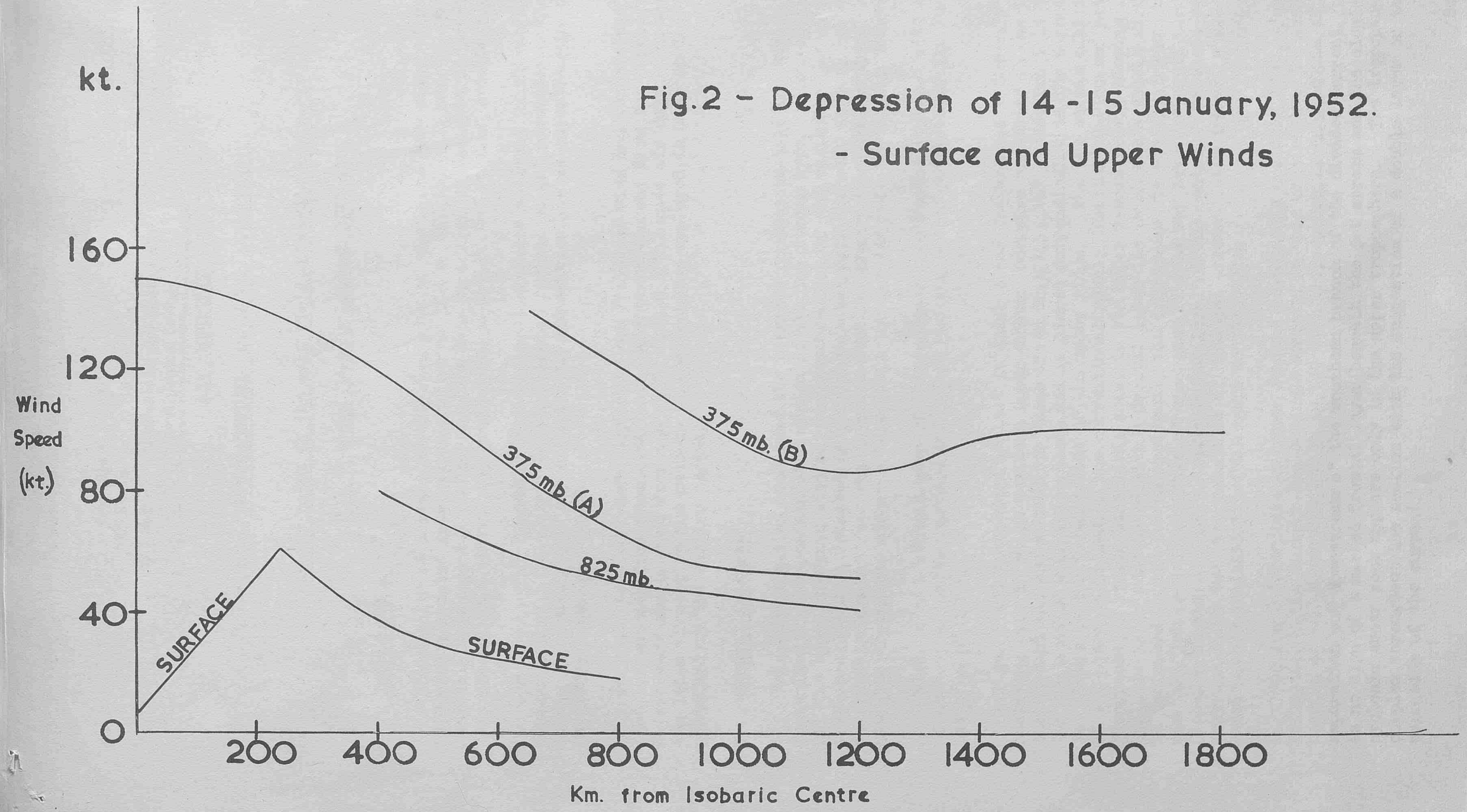


Fig.2 - Depression of 14 - 15 January, 1952.
- Surface and Upper Winds



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Further notes on the wind field of middle latitudes

by
D.H. Johnson (Meteorological Office)

1. Introduction

The results of a study of the structure of the upper westerlies of September 1950 were presented in a previous report: M.R.P. 667(1). It was thought desirable to subject a further month's data to a similar analysis in order to determine to what extent the conclusions of M.R.P. 667 were of general application.

The weather of September 1950 was predominantly westerly in character over western Europe. January 1950 was selected for the second examination since, during that month, the eastern Atlantic and western Europe experienced a variety of winter types of weather.

62 cross-section charts depicting the fields of wind and temperature along lines of section from Greenland to the Mediterranean and from the Azores to the Baltic were constructed from the daily 1400 G.M.T. upper air data. Details of the lines of section and methods of construction employed in both studies are contained in the earlier report.

2. General features

This investigation confirmed the previous conclusion that the upper westerlies always contain wind maxima of the jet stream type. More specifically, all the upper currents studied, whether branches of the westerlies or parts of large scale eddies detached or partly detached from the main stream, have essentially similar features in cross-section, namely a central maximum located near the tropopause within a baroclinic zone of the type previously described.

Figs. 1 to 4 contain some of the sections constructed together with copies of the corresponding surface and 300 mb working charts prepared in the Forecast Division of C.F.O.

Figs. 1 and 3 are each typical in their broad features of a commonly occurring type in which a strong upper current occupies one third to one half of the section and are similar to those given in M.R.P. 667. Section 1 is taken through a major warm anticyclone and is almost perpendicular to the northwesterly jet stream on the forward side of the upper ridge. The jet stream exhibits many of the characteristics of the conventional model; the maximum located near the tropopause, a strong baroclinic zone in the upper troposphere, a sloping zone of cyclonic shear associated with a frontal layer (in this case a double structure is indicated in the thermal pattern), the tropopause steeply sloping and broken near the jet centre, the axis of strongest flow at any level being to the right (looking downstream) of the axis at the level of the centre. Of these features only the first two were common to all the jet streams investigated, there being considerable variability with regard to the others.

Section 2 intersects an upper current immediately upstream of a major difluence. The current has a single maximum in this section but the abnormal spacing of the isokinetics high in the warm troposphere is symptomatic of the separation out downstream of the southern branch of the divided current. There is no sign of a marked frontal layer beneath the jet stream and the sloping cyclonic shear zone appears only in the upper troposphere. The tropopause is complex throughout the section with the suggestion of a double layer on the warm side of the current.

/Section

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Section 3 is taken through the central part of the jet stream illustrated in Fig. 2 after an interval of 24 hr. during which the jet stream system moved bodily southeastwards. The change in the shape of the isokinetics from Fig. 2 to Fig. 3 consists partly of a variation along the jet axis and partly of a variation in time. The soundings at Aldergrove, Liverpool and Larkhill give unusually clear evidence of a steeply sloping tropopause immediately to the north of the jet centre; frequently the evidence permits the analysis of either a continuous or discontinuous tropopause in this region.

Section 4 illustrates the principal features of those which intersected the large scale oscillation in the upper flow which affected Europe during the latter part of the month. The warm air thrust up over the northeast Atlantic and the Norwegian Sea is bounded to the north by a strong south-westerly current and to the south by a weaker easterly jet stream. Sloping across the British Isles is a broad quasi-isothermal zone which separates maritime from continental air whilst a shallow tongue of colder continental air has penetrated across western France and the Mediterranean. The warm air ascents indicate a double change of lapse in the region of the tropopause.

Over the area covered by the sections there were currents with central speeds in excess of 100 kt. on every day of the month. On 14 days there were jet streams of 140 kt. or more. The space distribution of the wind maxima found throughout the month along each section is contained in Figs. 5 and 6 which give the position and true speed (not merely the resolved component) of the core of each individual current. As previously noted (1) the heights of the maxima tend to increase with decreasing latitude. The frequency with which upper currents from directions between northwest through east to southeast occurred was about 20% as against 10% in the September study. The total number of maxima of all intensities was about the same in each month; 107 in January 1950 as against 104 in September. Figs. 7 and 8 contain frequency distributions of the levels at which the maxima occur. Over the British Isles the jet centres were usually found between 250 mb and 350 mb, a range 50 mb lower than that given for September 1950.

3. Observed velocity and shear profiles

The isobaric wind shears which occurred at the levels of the jet stream centres are illustrated by the velocity (component) and shear profiles given in Figs. 9 and 10. Fig. 9 consists of curves applicable to type C and type W* jet streams in addition to the overall mean profile. The profiles were constructed only from jet streams which flowed in a direction which was nearly normal to the lines of section. The shear profile was obtained as the first derivative of the overall mean profile given in Fig. 9 by estimating the slope at selected points.

The three velocity profiles are very similar in form to each other and to those obtained for September 1950. The speed at the centre is 124 kt as against 96 kt in September whilst the difference in speed between type C and type W jets is 26 kt in January as against 20 kt in September, type C jets being the stronger in both months.

It was noted in the September study that the average shear on the warm side of the current was about two thirds of that on the cold side. A similar ratio holds for the January mean profile, the speed falling off to half its maximum value in 300 and 180 miles on the warm and cold side respectively. From Fig. 10 it is apparent that the maximum shears on either side are also in this ratio. It is estimated that if the line of section had been varied so as to lie perfectly normal to the jet streams the values of the mean speeds and shears would be 5 to 10 per cent greater than those given in Figs. 9 and 10.

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*Referred to as "warm" and "cold" jets in M.R.P. 667 but subsequently defined as "type C" and "type W" in "Structure of the upper Westerlies" published in Q.J. Roy. Met. Soc. Type C jets are those which tend to move laterally towards the warmer tropospheric air and type W jets are those which move towards the colder tropospheric air.

The individual profiles from which the mean profiles were constructed were all of the same general form as the mean (cases of double maxima being excluded), although there was considerable variation in the relative intensities of the anticyclonic and cyclonic shears. However, in no case did the anticyclonic shear exceed the cyclonic shear as it does in some profiles for 80°W published by Staff Members, University of Chicago (2) and Palmén (3).

The greatest observed cyclonic shear was again of the order of 100 kt. in 100 miles, 100 miles being a lower limit to the distance over which it is possible to estimate isobaric shear with this network. A recent experiment in photographic reconnaissance by Hurst (4) suggests that shears of up to twice this value may occur over shorter distances. It is of some interest that Hurst's data for 18th January 1952 gives a velocity profile very similar in general characteristics to the mean profile obtained in this analysis.

4. Vertical wind shears

The remarks of paragraph 4(b) of M.R.P. 667 were found to apply to the month of January 1950 without modification.

5. Association between jet streams and surface fronts

The relationships between jet streams and surface fronts were found to be as diverse in January as in September. Figs. 1 to 4 contain examples. On some occasions the general pattern of the upper flow could be inferred from the surface configuration, on others this was impossible.

6. The mean January flow from Greenland to the Mediterranean

Fig. 11 shows a mean cross-section from Angmagssalik to Tunis for January 1950. Mean values of temperature and normal wind component were computed from the individual sections over a network of points at vertical and horizontal separations of 3000 ft. and 150 miles.

In Fig. 12 are tephigram plots of mean "ascents" at the points marked A, B, C and D in Fig. 11. It is not possible to represent the tropopause on the mean section with entire satisfaction. On the cold side of the mean current the level of transition from mean tropopause to mean stratosphere can be placed within fairly narrow limits. On the warm side however, the placing of a mean tropopause from the mean ascent curves becomes more arbitrary. Following Hess (5) and Hutchings (6) the mean warm air tropopause has been placed at the level of minimum temperature although it is clear from the mean ascent curves that this level will not correspond with the statistical means of the individual tropopause heights.

(a) The N. Atlantic winter jet.

The mean wind field has a maximum of 60 kt at a height of about 9 km in latitude 61N. This current, although of lesser intensity, is broadly similar to many individual jet streams and is located in a position favoured by the strong upper westerlies on many days of the month (see Fig. 5).

The latitude of the mean current is higher than those of wind maxima found on mean sections published for other longitudes of the N. Hemisphere. Hess, in his winter section for 80°W (5) found a maximum of 50 m.p.s. (100 kt) at a height of 12 km in latitude 35N. To obtain some idea of the representativeness of the section for January 1950 a corresponding mean winter section was constructed from the mean flow charts given by Brooks et alia (7). This is reproduced in Fig. 13 with the mean 300 mb contour chart. The scale of the mean flow charts as published is too small for the pattern of Fig. 13 to be correct in detail but the broad features are given with sufficient accuracy. Comparison of Fig. 11 with Fig. 13 indicates that the mean flow pattern for January 1950 over the eastern N. Atlantic was quite typical of the winter season and that it is normal for a mean south westerly winter current to be found in rather high latitudes. It is well known that the 80°W meridian lies near the axis of a quasi-permanent cold trough; the Atlantic section is taken across a mean warm ridge.

/The

The height difference between the 80°W and E.Atlantic mean currents is a reflexion in the mean of the tendency for the individual currents to flow at higher levels in low than in high latitudes. The greater mean speed of the 80°W current is presumably due in part to a greater relative strength of individual 80°W jet streams (the strongest upper flow is often concentrated to the south of an upper trough) and in part to the lesser variability of the flow over eastern N. America noted by James (8). The upper flow apparently enjoys a greater degree of lateral freedom over the east Atlantic.

(b) Stratospheric Currents

To the north of a (westerly) jet stream there occurs in the stratosphere a band of relatively warm air and further north than this in mid-winter the winds are often found to increase with height in the stratosphere. It is not surprising therefore, that in the northern most part of the mean January 1950 section, the wind increases with height both below and above the tropopause. Occasionally when a jet stream was located in high latitudes the wind reports from Angmagssalik suggested the presence of a separate maximum in the arctic stratosphere. (See example in Section 4). It seems that on occasions a high speed stratospheric current occurs in association with the strong thermal gradient which must exist between the warm band to the north of a high latitude jet and the intensely cold arctic stratosphere noted by Palmén (3). The effect could not be studied in detail since it occurred at the extreme edge of the section.

(c) Low latitude jet streams

The evidence in the mean for a high level low latitude maximum is of interest in view of Palmén's suggestion that there exists a quasi-permanent subtropical jet stream as distinct from the more mobile westerly jets of higher latitudes (9). An examination of the daily sections in conjunction with high level contour charts showed however that the Mediterranean currents were much more variable in position and intensity than the quasi-permanent winter jet stream which flows south of the Himalayas (described by Chaudhury (10) and Yeh (11)) and into which they presumably feed. On most occasions when a distinct upper current appeared in the extreme south of a section it clearly originated in higher latitudes, either branching from the higher latitude westerlies over the N. Atlantic or forming the lower limb of some large scale oscillation of the upper flow over western Europe such as that illustrated in Fig. 4.

Lack of data at this period prevented the extension of our sections into lowest latitudes and in the absence of such data no definite conclusion as to the existence south of the Mediterranean and westwards of a quasi-permanent current could be reached.

7. Future research

M.R.P. 667 concluded with suggestions of particular problems which deserve special attention. The present study indicates the need for an examination of data from very high and low latitudes in the N.Atlantic and W.European sector.

The conclusions of the earlier study having been confirmed it seems likely that future work would most profitably take the form of analysis of specific properties of the upper flow and studies of individual case histories.

/References

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(10)	Chaudhury, A.M.	1950	Tellus, Vol.2, p.56.
(11)	Yeh, T.C.	1950	Tellus, Vol.2, p.173.

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CONTOURS OF THE 1000mb AND 300mb SURFACES ON 3rd. JANUARY 1950

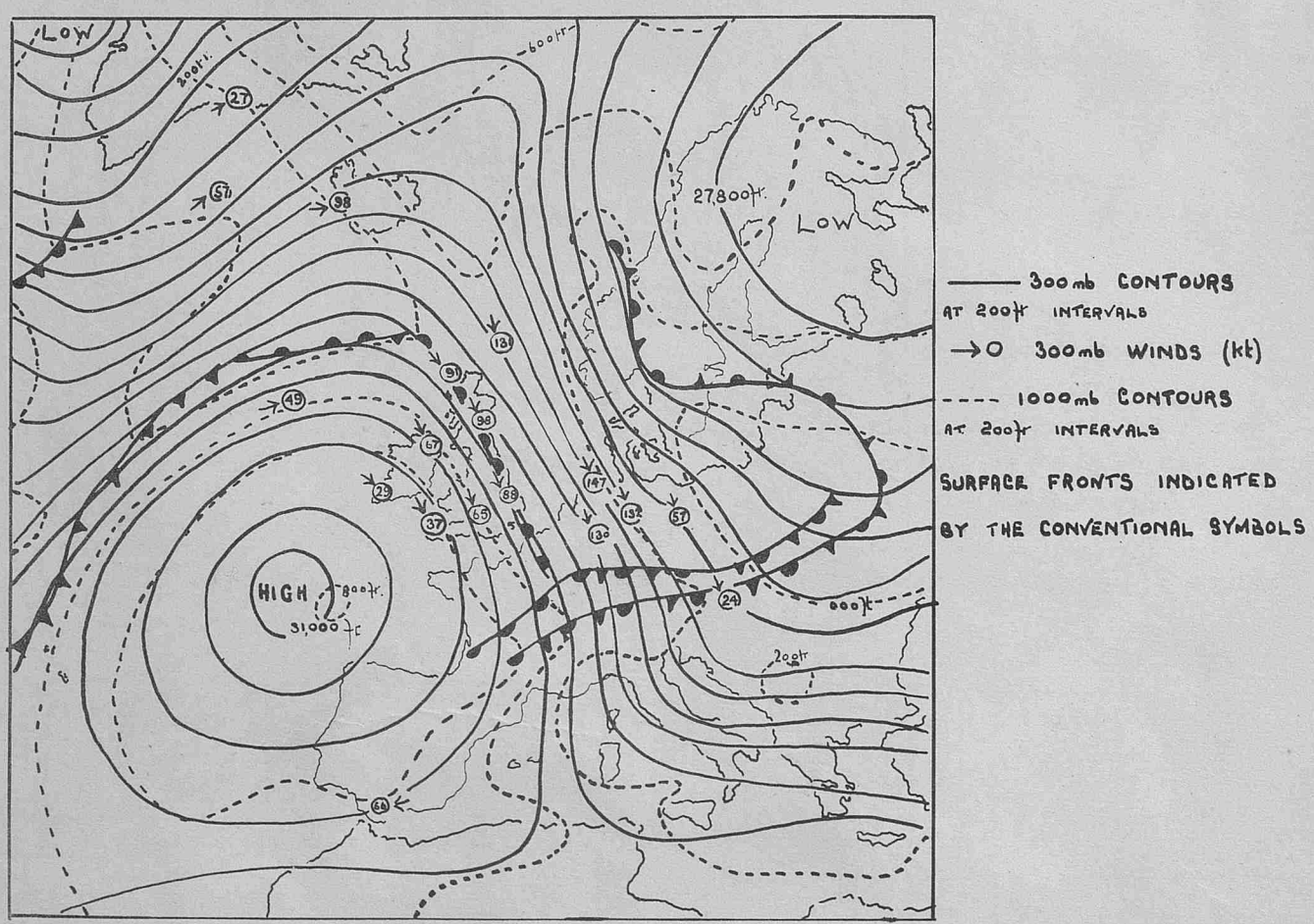
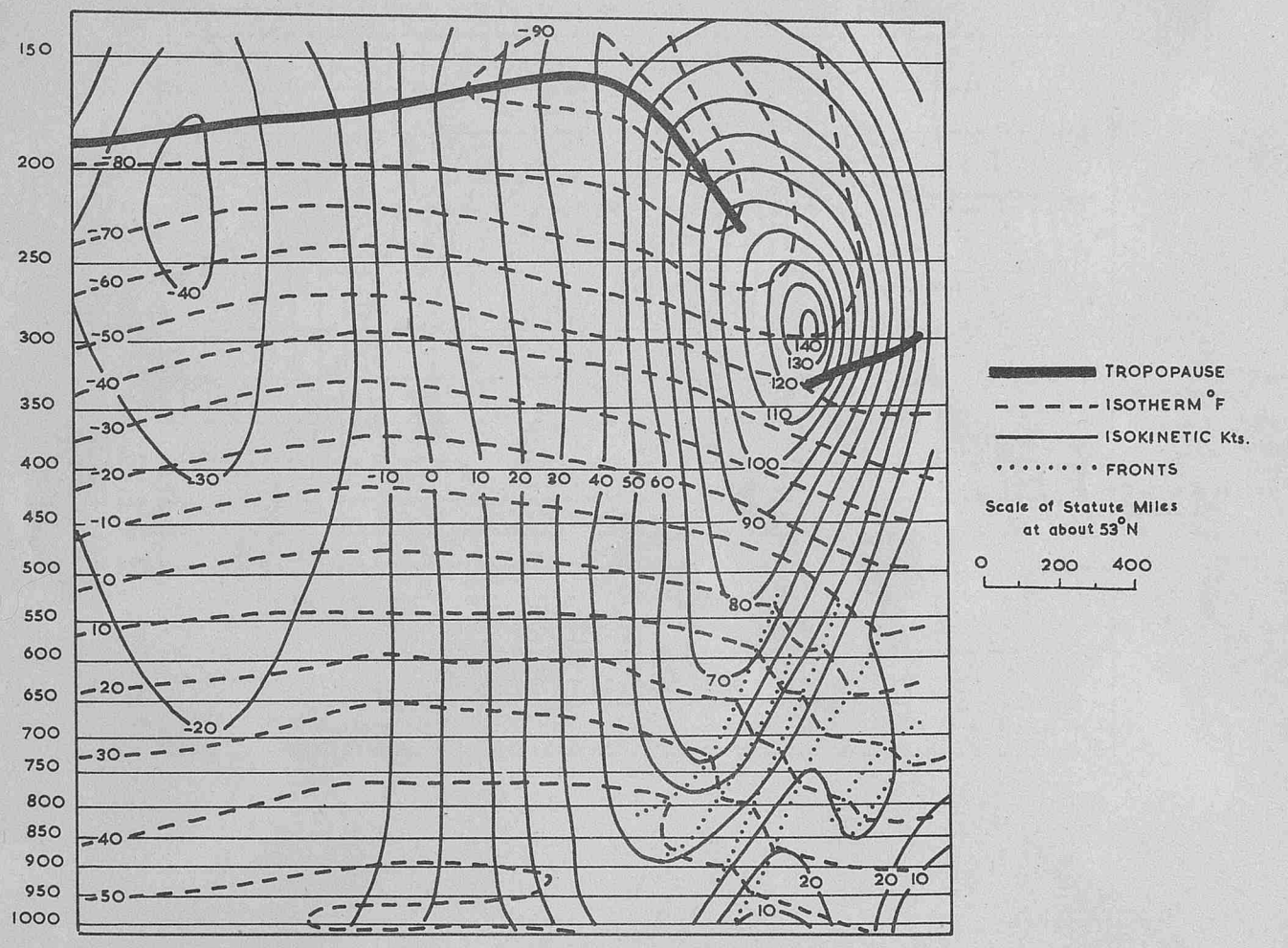


FIG I. SECTION I. CROSS-SECTION FROM LAGENS – COPENHAGEN. 1400 G.M.T. 3 JANUARY 1950



CONTOURS OF THE 1000mb AND 300mb SURFACES ON 3rd. JANUARY 1950

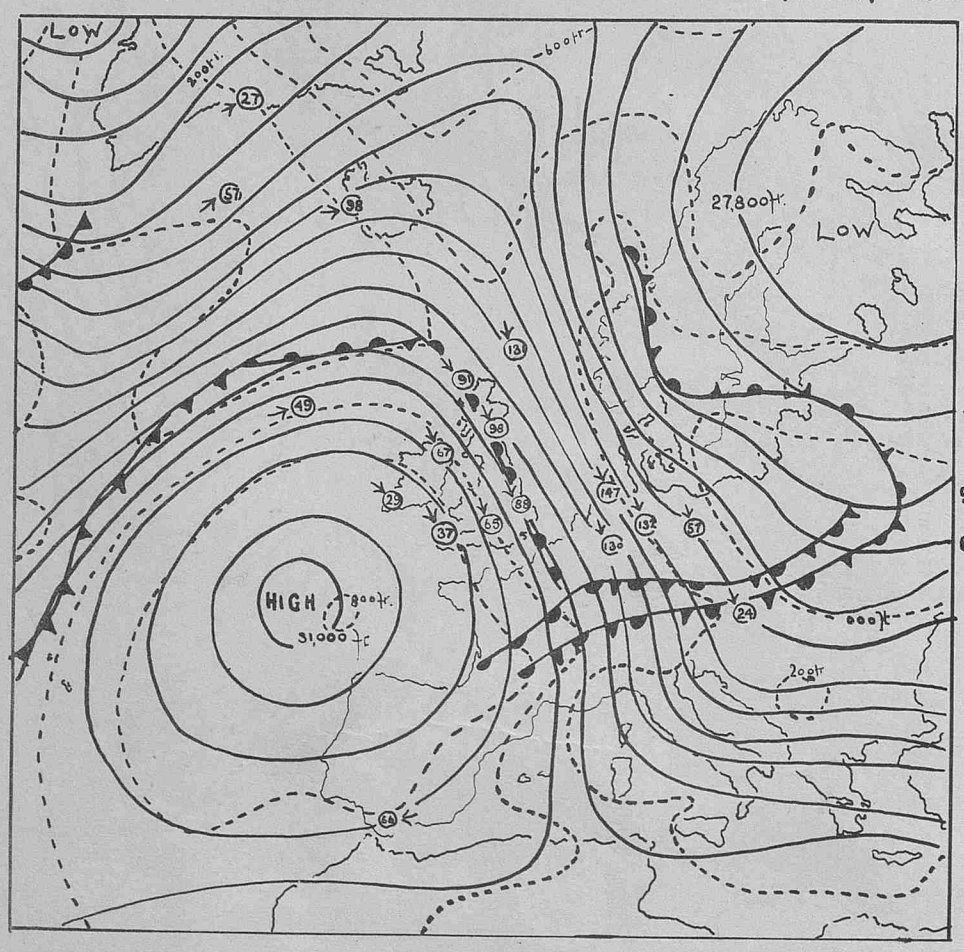
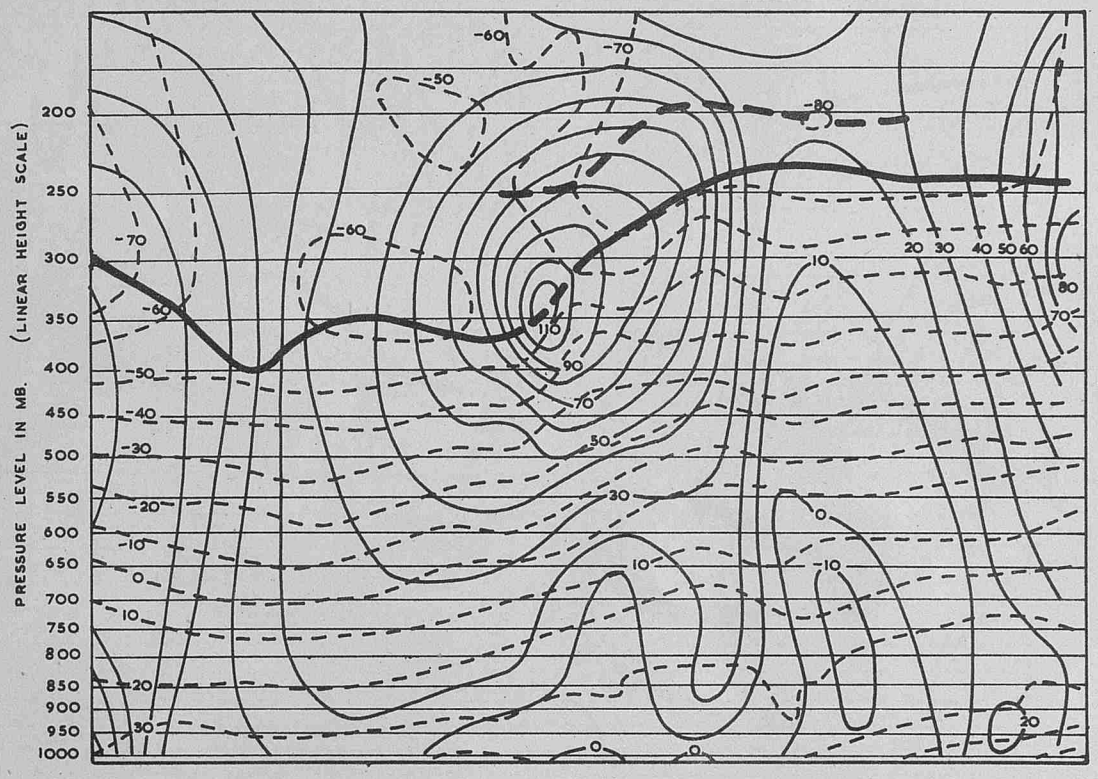


FIG 2 SECION 2 CROSS SECTION FROM ANGMAGSSALIK – MALTA 30 JANUARY 1950



CONTOURS of the 1000 mb. and 300 mb SURFACES at 1500 G.M.T. on 30rd. JANUARY 1950

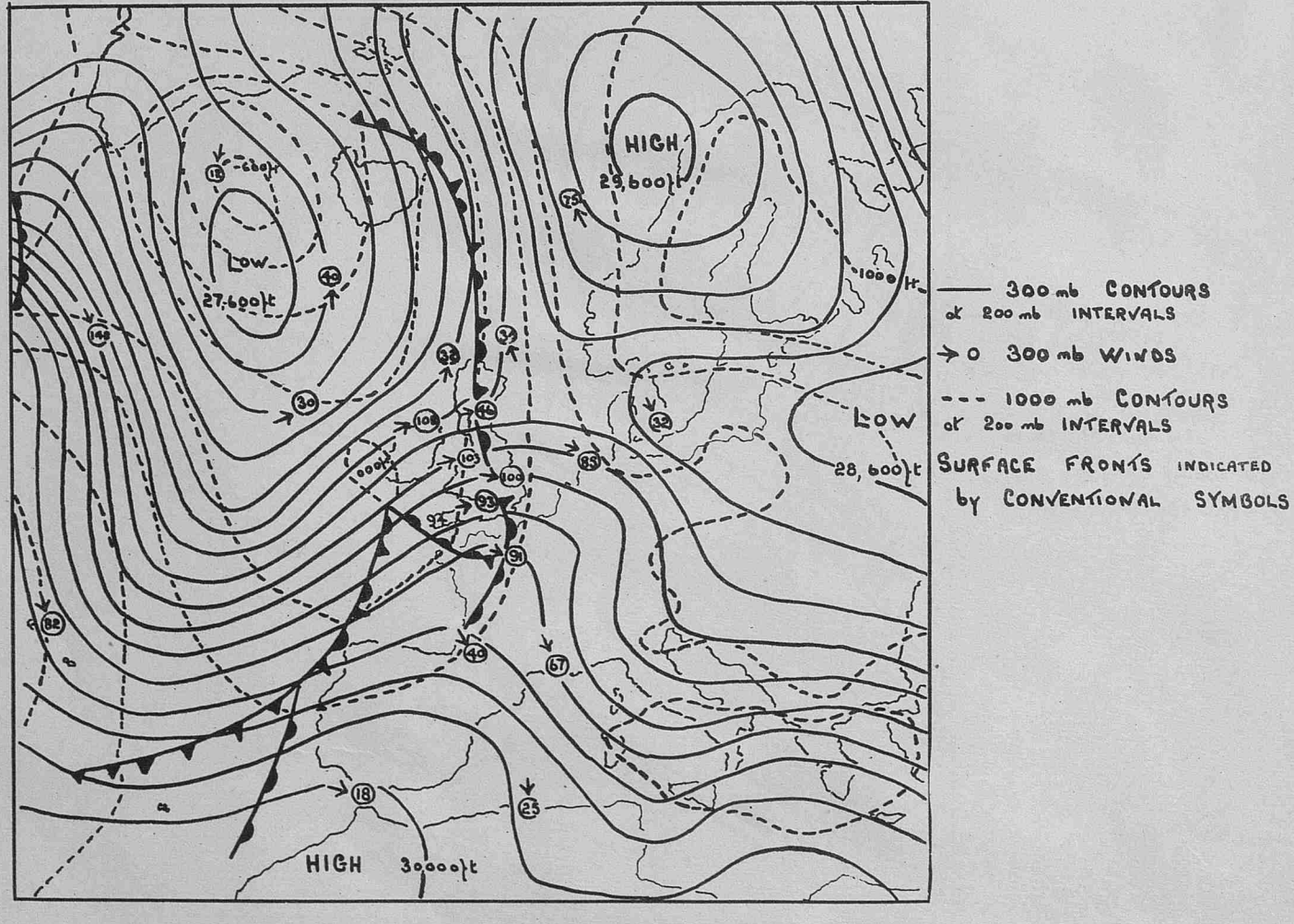
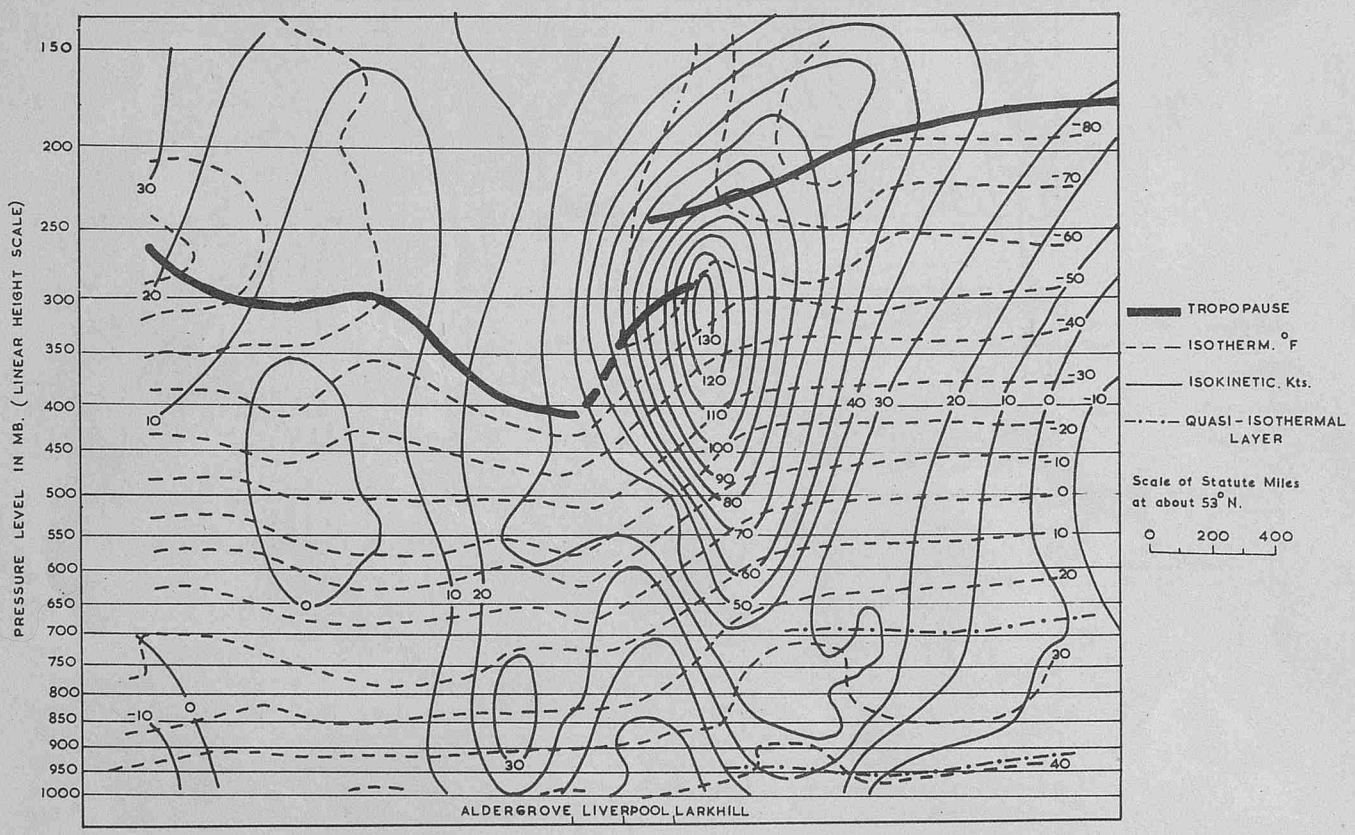


FIG 3. SECTION 3 CROSS-SECTION FROM ANGMAGSSALIK – MALTA 1400 G.M.T. 31 JANUARY 1950



CONTOURS of the 1000 mb and 300 mb SURFACES at 1500 G.M.T. on 31st. JANUARY 1950

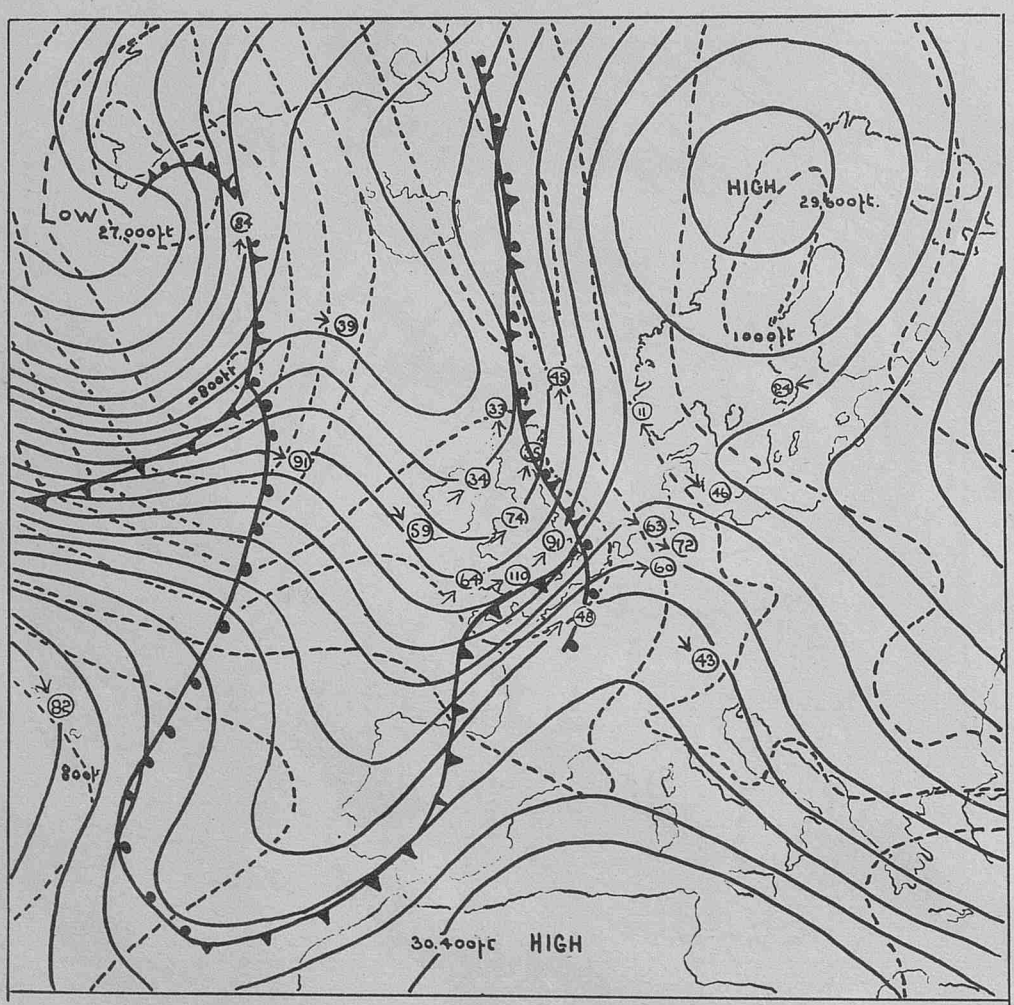
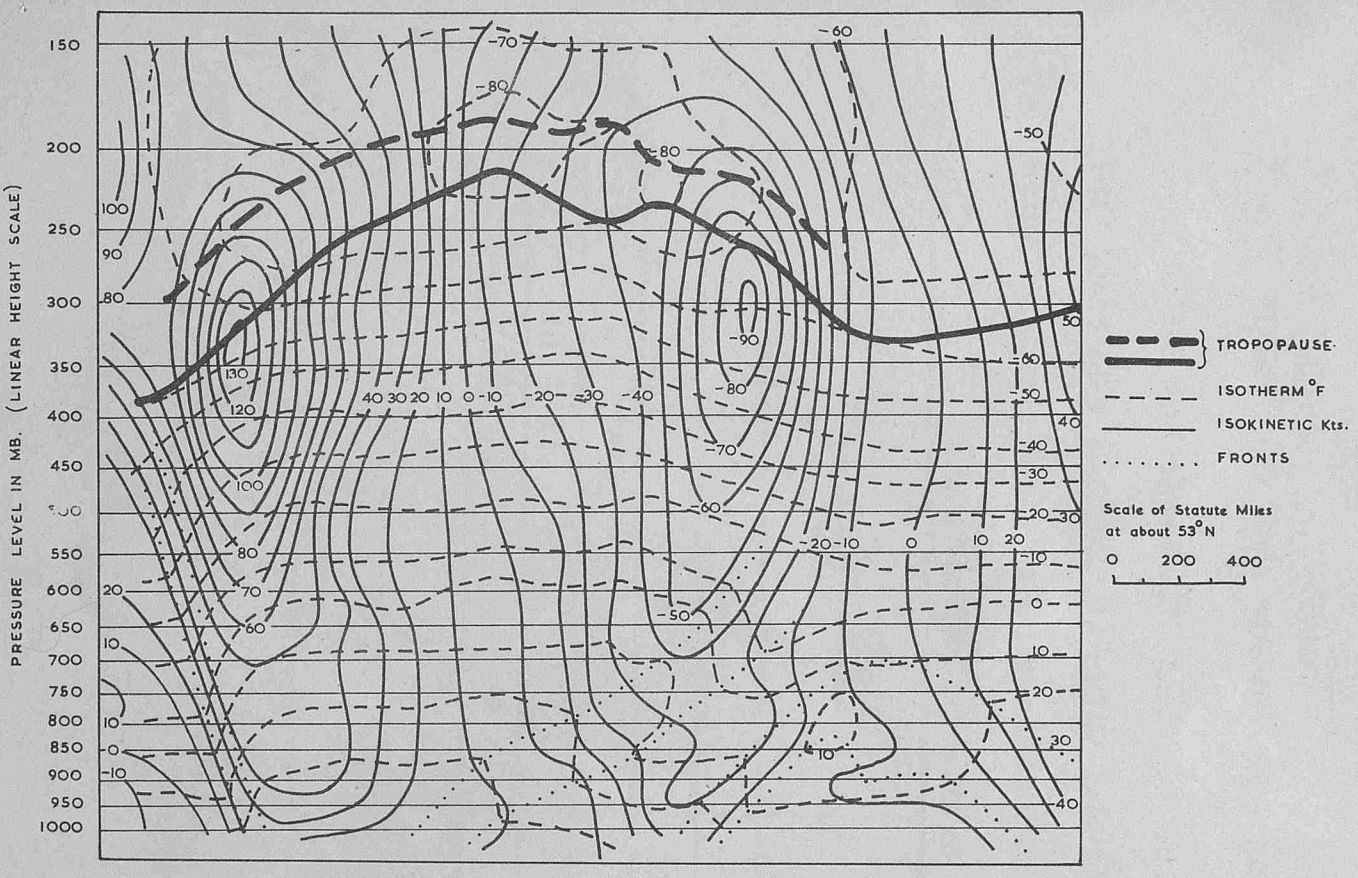
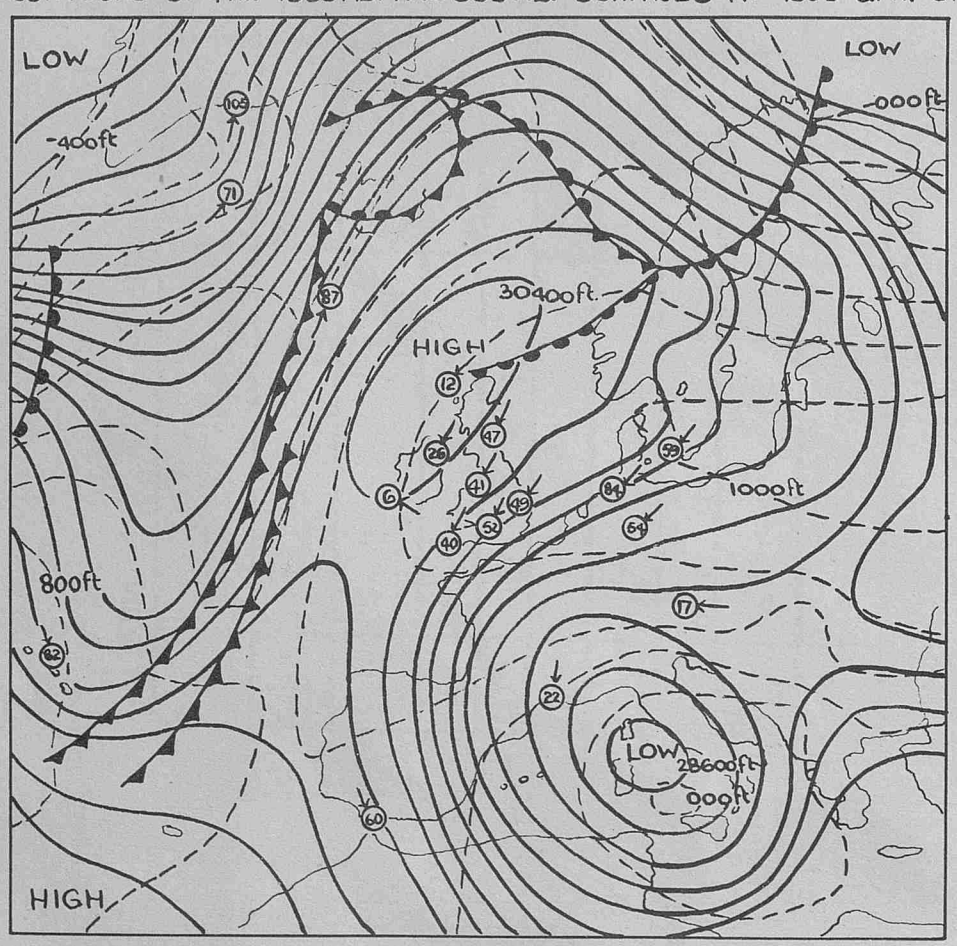


FIG 4 SECTION 4 CROSS-SECTION FROM ANGMAGSSALIK - TUNIS 1400 G.M.T. 19. JANUARY 1950.



CONTOURS OF THE 1000 MB. AND 300 MB. SURFACES AT 1500 GMT. ON 19TH JANUARY 1950.



300 mb. CONTOURS AT 200 ft INTERVALS.
300 mb. WINDS Kt.
1000 mb. CONTOURS AT 200 ft INTERVALS.
SURFACE FRONTS INDICATED BY CONVENTIONAL SYMBOLS.

FIG. 5. POSITIONS AND MAGNITUDES (AT 1400 GMT) OF CENTRES OF MAXIMUM VELOCITY IN AIRSTREAMS CROSSING THE SECTION ANGMAGSSALIK–MALTA IN JANUARY 1950.

The figures give the velocity at the centres of the maxima in tens of knots (positive for north easterly flow.) The figures in brackets refer to currents crossing the section to the south of Malta and give the velocity of the currents at the end of the section (not at their centres.)

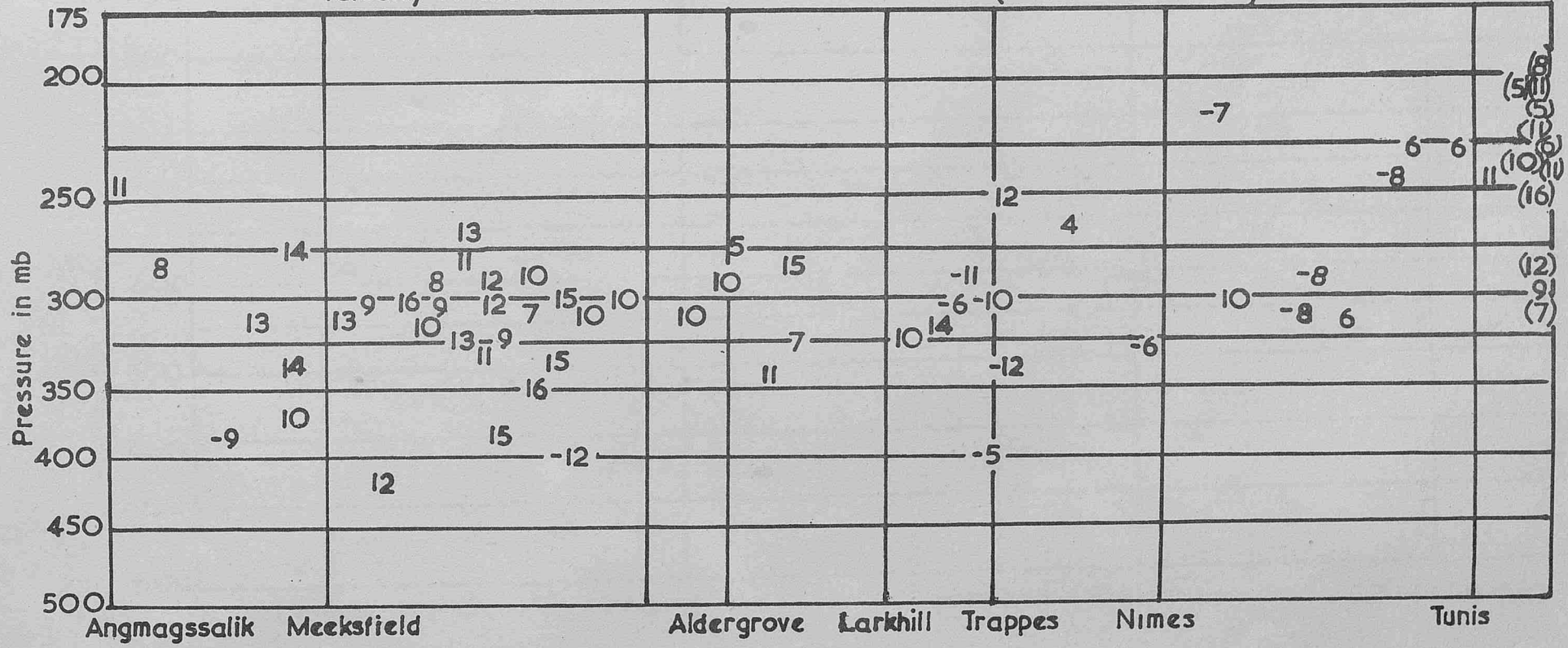
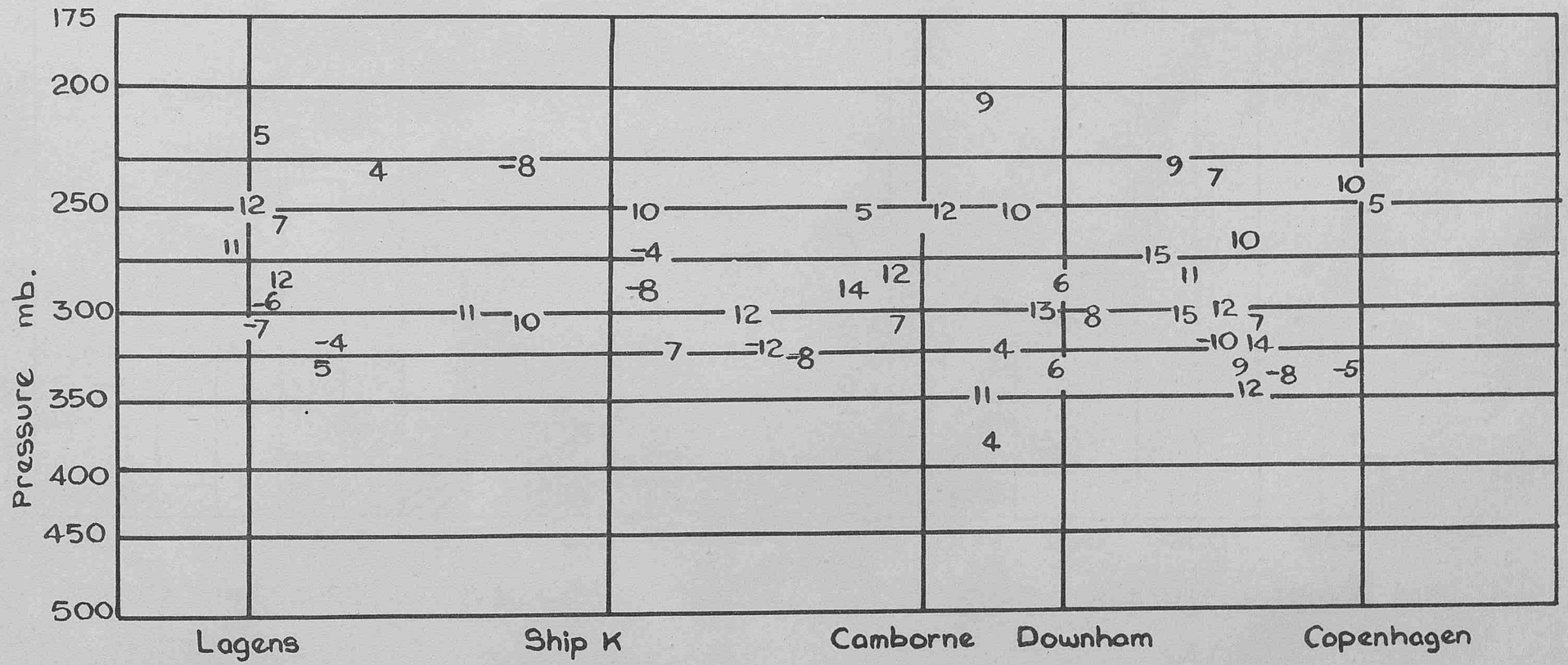


FIG. 6. POSITIONS AND MAGNITUDES (AT 1400 GMT.) OF CENTRES OF MAXIMUM VELOCITY IN AIRSTREAMS CROSSING THE SECTION LAGENS-COPENHAGEN IN JANUARY 1950.

The figures give the velocity at the centres of the maxima in tens of knots (positive for north-westerly flow.)



FREQUENCY DIAGRAM OF THE WIND MAXIMA FOR JANUARY 1950.

FIG. 7

SECTION ANGMAGSSALIK–MALTA

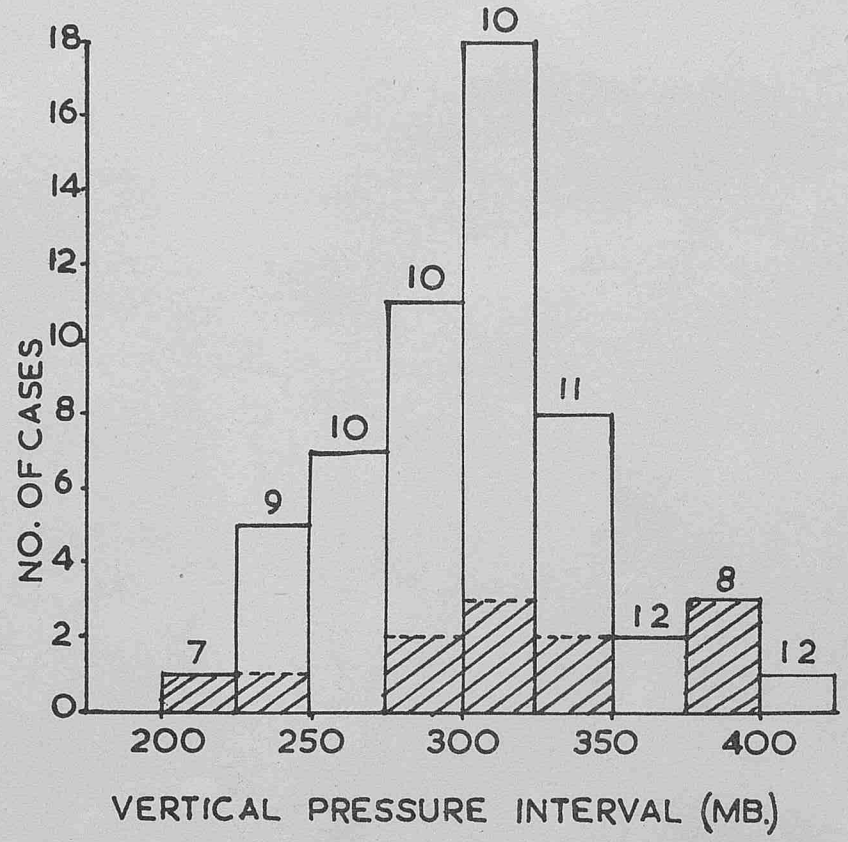
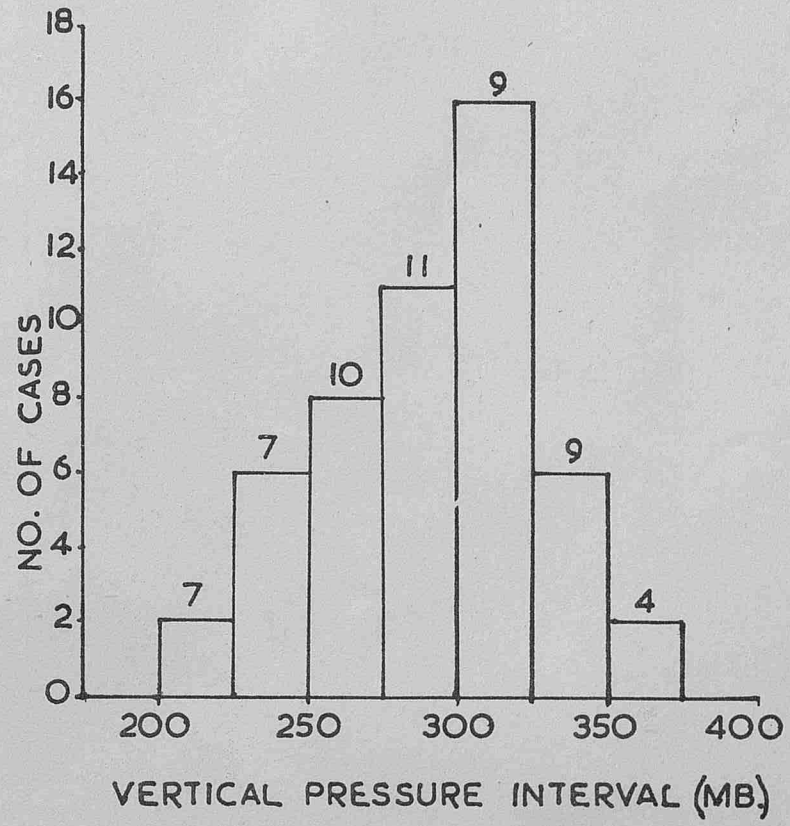


FIG. 8

SECTION LAGENS–STOCKHOLM.



FIGURES ABOVE THE HISTOGRAM GIVE THE MEAN VALUE OF THE WIND MAXIMA (TENS OF KNOTS) OCCURRING IN THE GIVEN PRESSURE INTERVAL. HATCHED AREAS REFER TO MAXIMA WITH A NORTHEASTERLY COMPONENT.

FIG. 9. MEAN VELOCITY PROFILES AT THE LEVELS OF THE CENTRES OF JET STREAMS

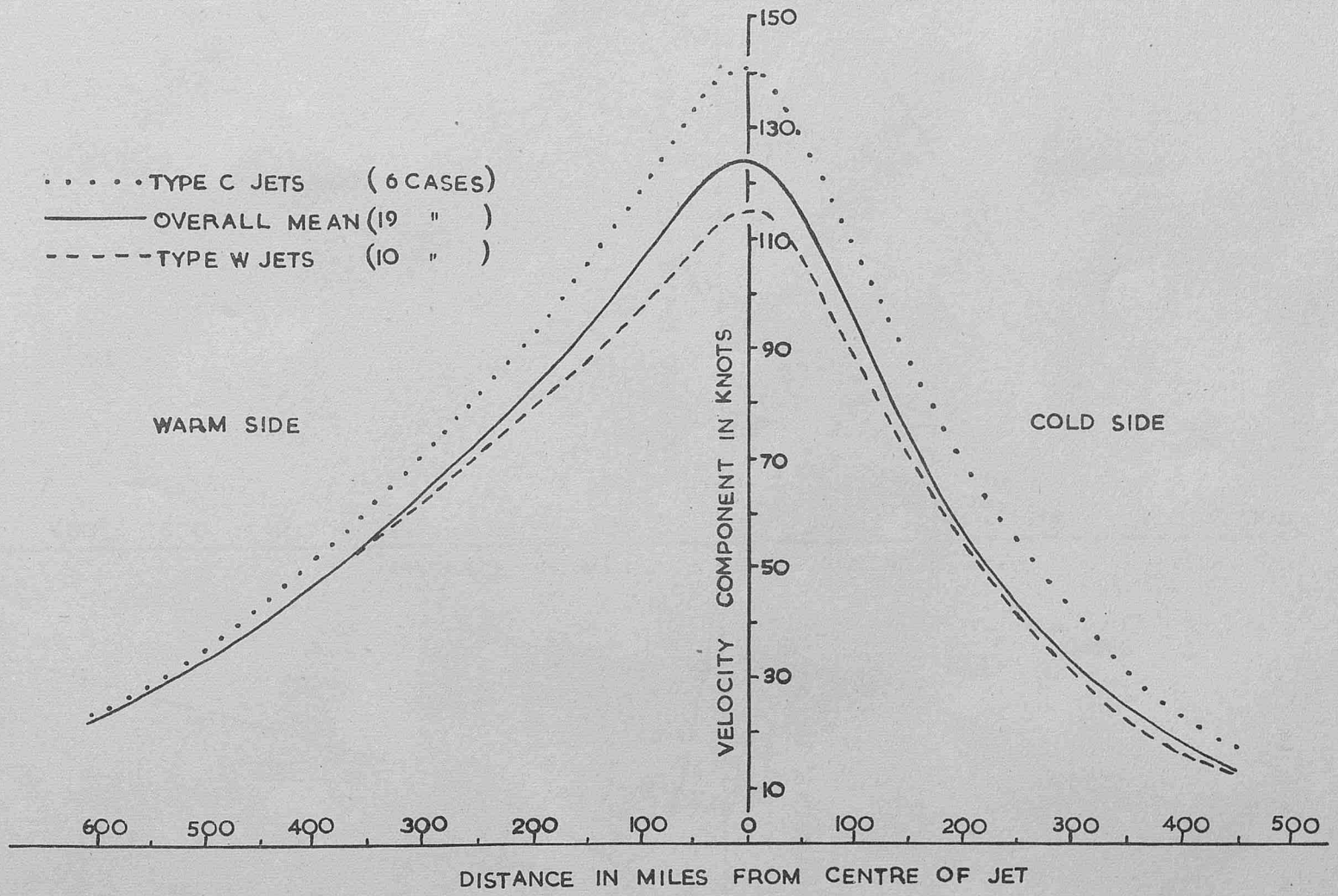


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FIG 10.

MEAN ISOBARIC SHEAR PROFILE AT THE LEVELS OF THE CENTRES OF JET STREAMS

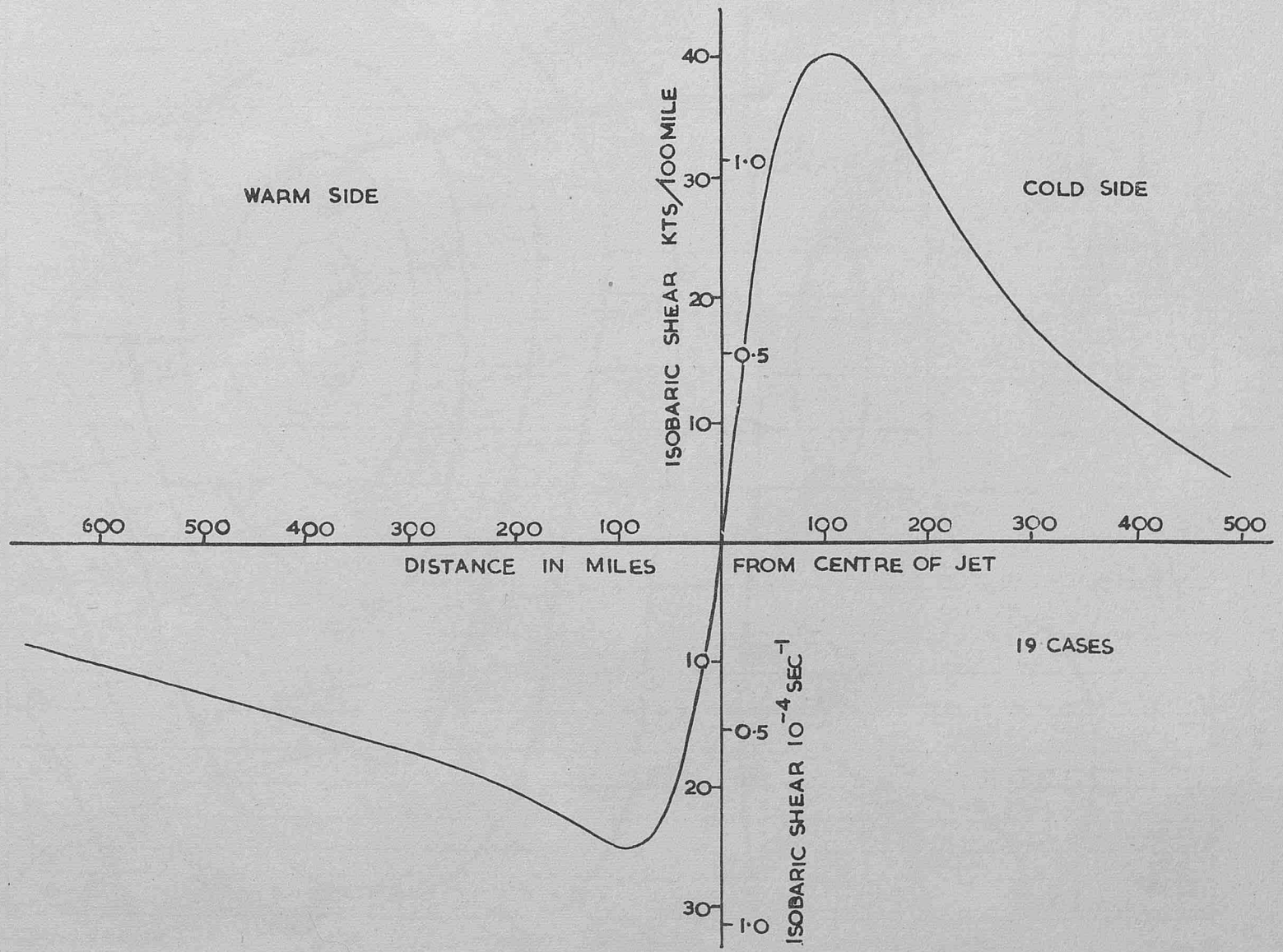
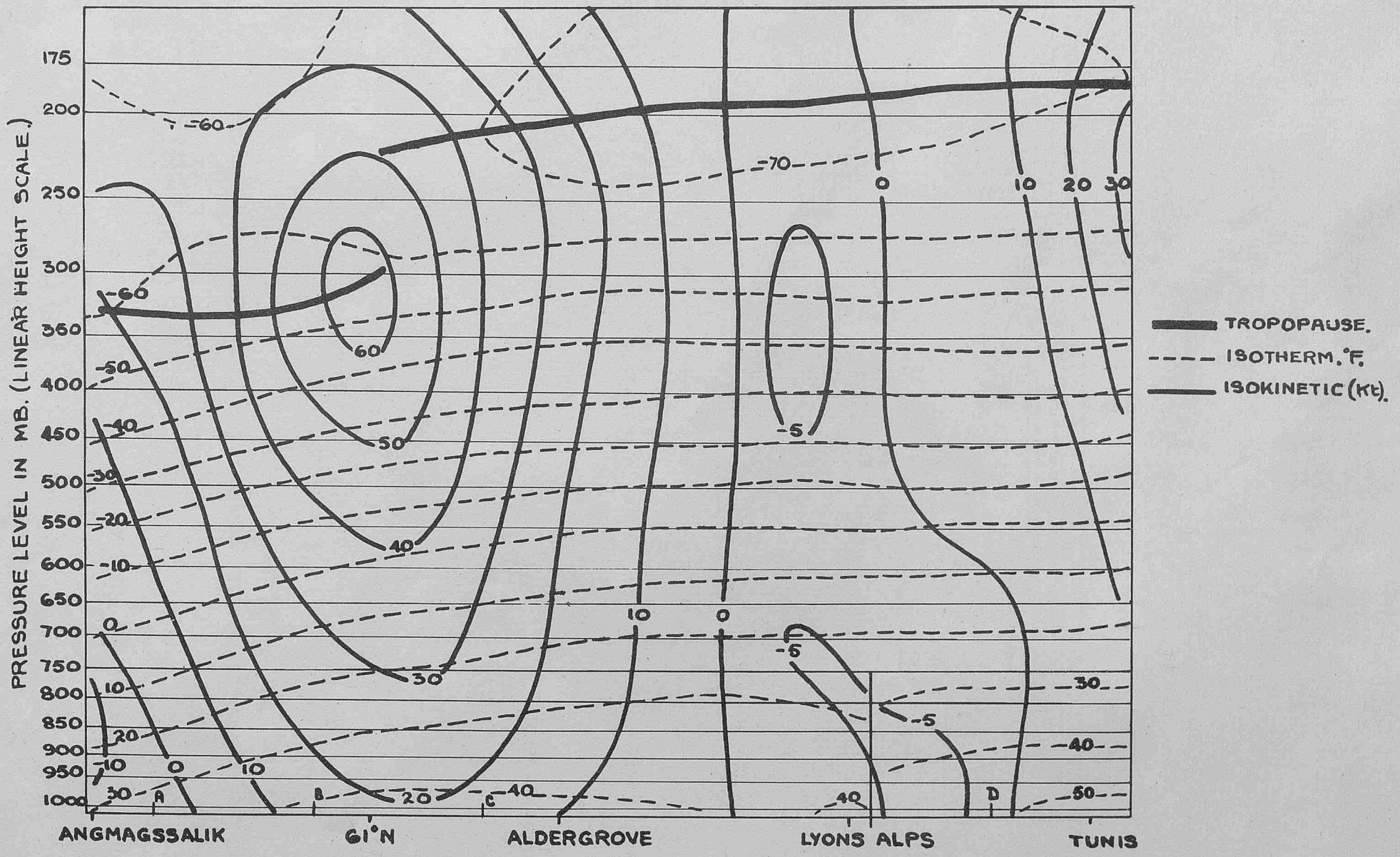


FIG 11 MEAN CROSS-SECTION FROM ANGMAGSSALIK TO TUNIS FOR JANUARY 1950 (1400 GMT.)

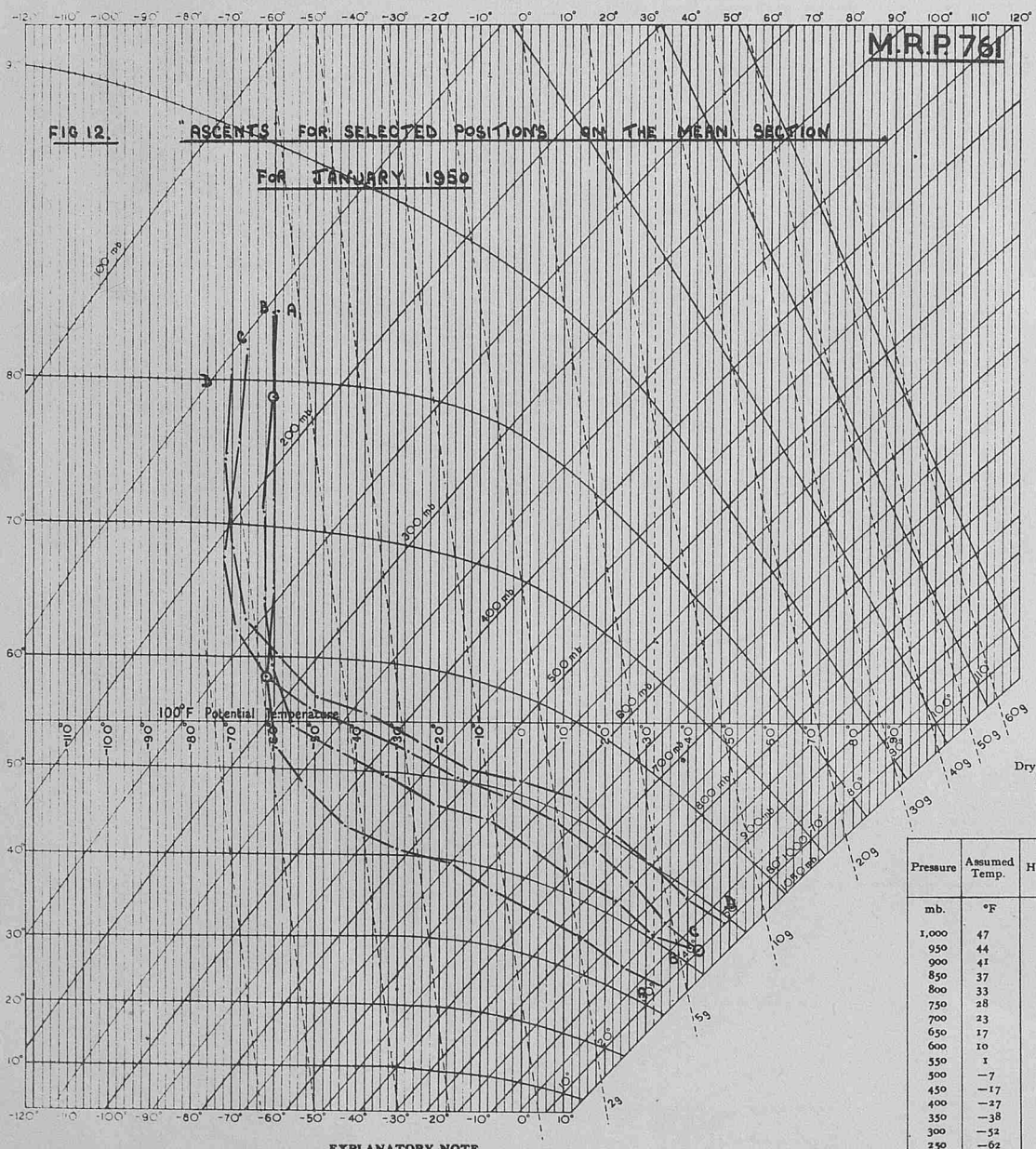


Form 2811 AIR MINISTRY, METEOROLOGICAL OFFICE DATE 19

TEPHIGRAM OF RADIO-SONDE RECORDS

STATIONS	TIMES

REMARKS



Dry Adiabatic Lapse Rate is 5·3°F. per 1,000 feet.

Pressure	Assumed Temp.	H. Height above 1,000 mb.	Increase of height due to 10°F increase in mean temperature.
mb.	°F	Feet.	Feet.
1,000	47	0	—
950	44	1,380	30
900	41	2,830	60
850	37	4,350	90
800	33	5,950	120
750	28	7,640	150
700	23	9,420	190
650	17	11,320	230
600	10	13,340	270
550	1	15,490	320
500	−7	17,810	370
450	−17	20,330	430
400	−27	23,080	490
350	−38	26,120	560
300	−52	29,530	640
250	−62	33,440	740
200	−66	38,150	860
150	−66	44,200	1,000
100	−66	52,700	1,200

Adjust heights to other surface pressures by adding (subtracting) 28 feet for every millibar of pressure above (below) 1,000 mb.

EXPLANATORY NOTE

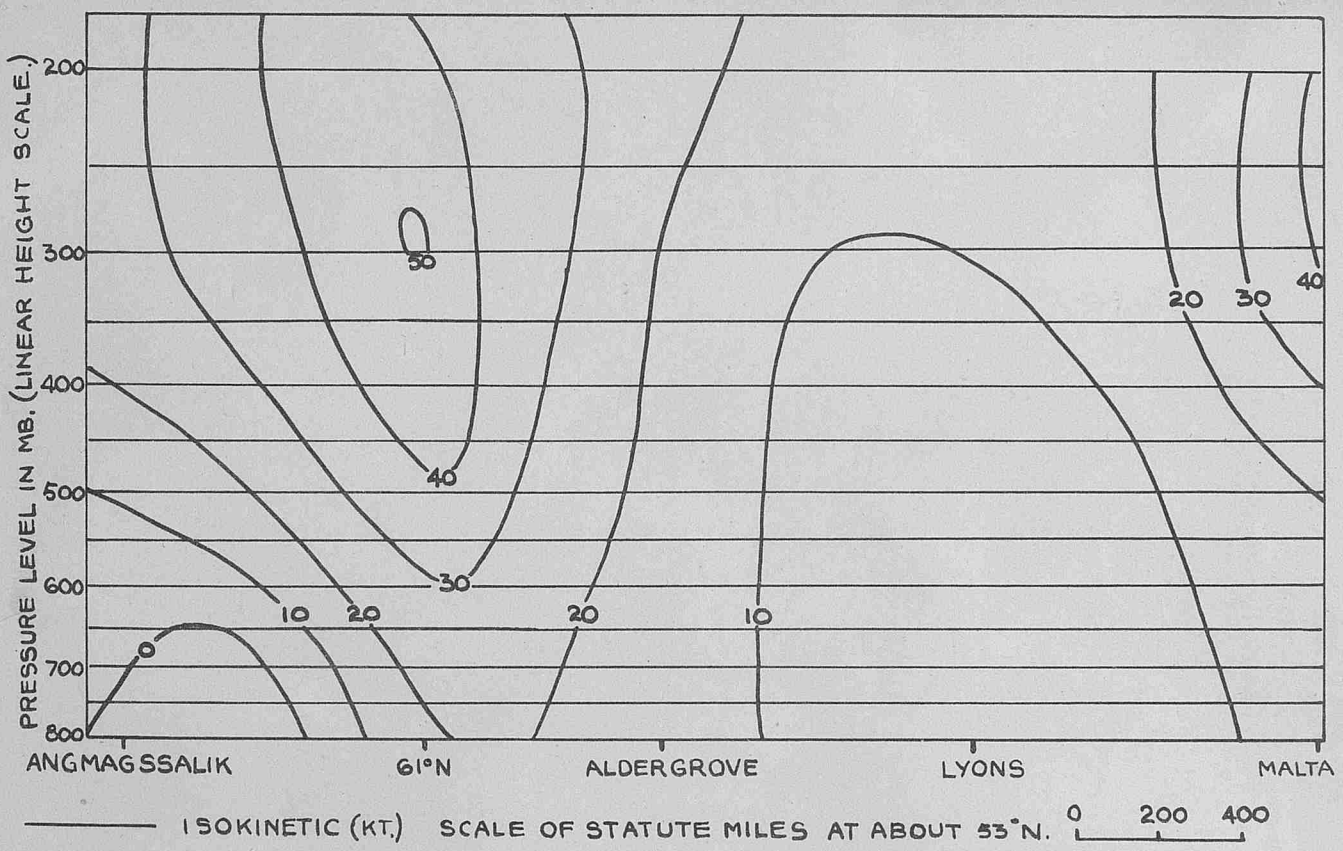
This chart is intended for use with radio-sonde ascents to show the general features of the complete results at a glance. It is not intended for use where the main purpose of plotting the result is to examine the details in the lower part of the troposphere. The scale of the form is approximately half that of the Tephigram Form 2810.

The temperature figures printed along the 1,000 mb. line apply both to the saturated adiabatics (the curved lines) and to the temperature scales (the vertical lines). The values printed at the left hand side of the diagram apply only to the saturated adiabatics.

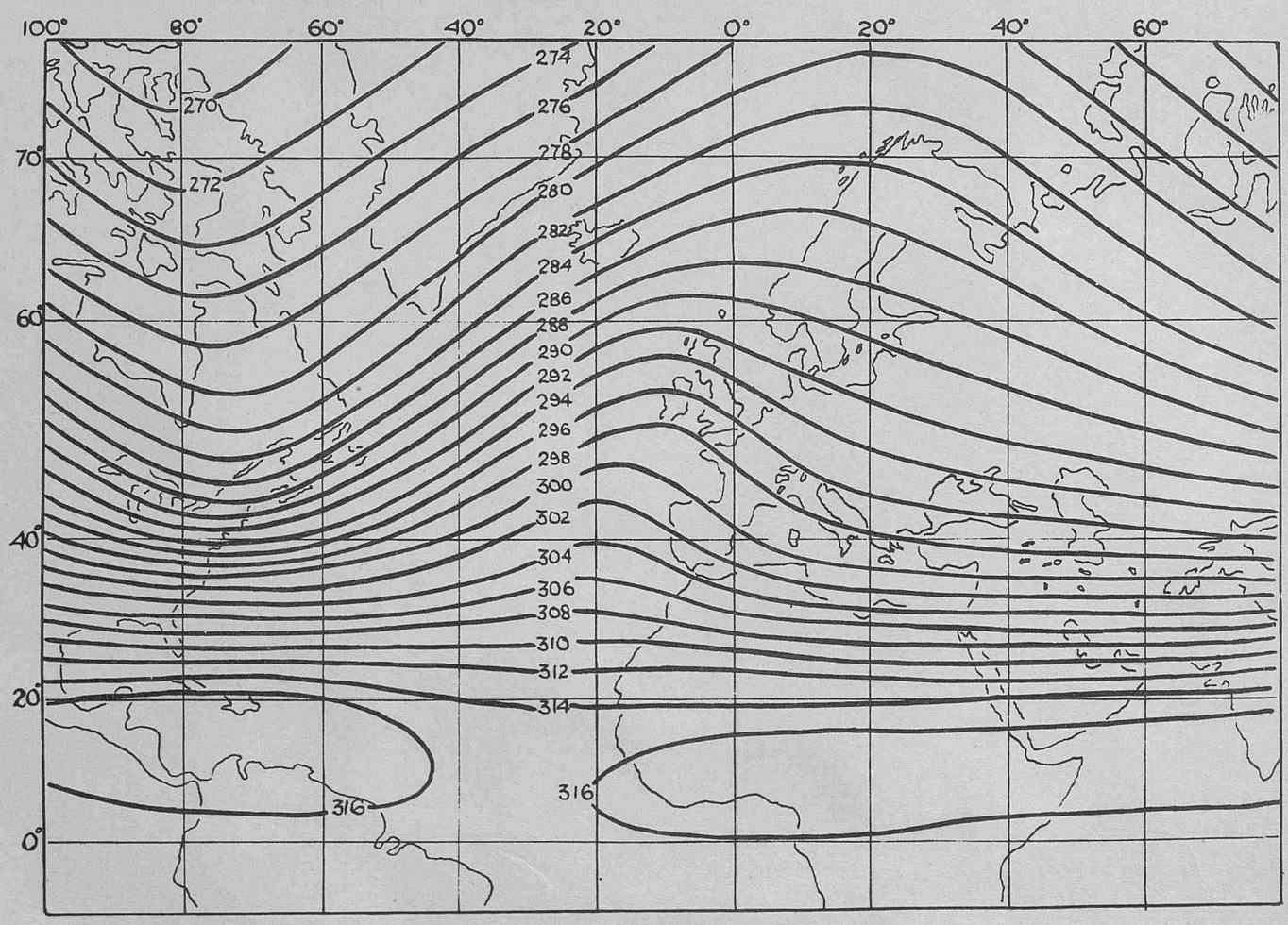
A straight line on the diagram gives a constant lapse rate. For an inversion the curve must cross the temperature scale from left to right. If the result of a radio sonde ascent is plotted it is usually easy to see where the stratosphere begins. It may begin as low as 400 mb.

(F.1662). Wt.46346 20,000. 3,43. Gp.959. FOSH & CROSS LTD.

FIG. 13. MEAN WINTER CROSS-SECTION FROM ANGMAGSSALIK TO MALTA.



MEAN 300MB CONTOURS FOR DECEMBER TO FEBRUARY



At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No ..... by Mr .....
A copy is available in the Library of the Meteorological Office.

M.R.P. 762
S.C. II/124
30th October, 1952.

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METEOROLOGICAL RESEARCH COMMITTEE

Rainfall of depressions which moved eastward across the British Isles between the Bristol Channel and the Scottish Border.

by
J.S. Sawyer (Meteorological Office)

1. Introduction

In a previous investigation (M.R.P. 748) a study was made of the rainfall distribution over the British Isles resulting from depressions which passed eastward between Scotland and the Faeroes. The present paper reports the results of a similar study of the rainfall of depressions which moved eastward across England.

2. Method

The methods adopted were essentially the same as those adopted in the first investigation.

By examination of the Daily and Monthly Weather Reports all occasions were noted on which a depression moved eastward across England north of the Bristol Channel during the 10 years, 1941 to 1950. In order to provide a homogeneous set of data it was required that there should be a closed isobar on at least one of the charts published in the D.W.R., and that the depression centre should remain south of the Scottish Border and north of a line joining the Bristol Channel to the Thames estuary throughout its passage across the country. 45 depressions were found during the period which satisfied these conditions.

For each depression a rainfall chart was drawn for the British Isles on the basis of observations printed in the Daily Weather Report. The period covered by the chart was chosen to include all of the frontal rain from the depression; showers in the adjacent air masses were necessarily included, and difficulty in separating the rain from that of preceding or following systems arose in some cases. However no serious doubt arose regarding the general shape and intensity of the rainfall distribution. Most of the rainfall charts covered a period of 36 hours or less, but in some cases it was necessary to include a longer period; for one depression which became slow moving in the North Sea this extended to 84 hours.

For each depression a 1000-500 mb thickness chart was prepared and examined; this was drawn for a time so chosen that the chart should represent the thermal structure of the depression when it crossed England. Observations were taken from the German Daily Weather Reports to complete the British charts for the war-period.

3. The average rainfall distribution

Fig. 1 shows the average rainfall distribution for the 45 depressions which moved eastward across England. It is based on the average rainfall values shown on it and, in calculating these, values have been interpolated from the rainfall charts when observations were not available in the Daily Weather Report. Since the stations for which observations were available in the D.W.R. are nearly all on relatively low ground the chart does not adequately represent the rainfall of the hilly areas particularly in Wales and the Pennines where orographic effects are strongly developed. In order to provide some indication of the magnitude of the rainfall in such notably wet areas, the average rainfall for the same

/depressions

depressions was calculated from the autographic records of Cwm Dyli (near Snowdon) and of Cray Reservoir (Breconshire). These are inset on the chart together with the value for Eskdalemuir, also a high level station, taken from the Daily Weather Report.

Fig. 1 shows that the average rainfall amount from these depressions is fairly uniform over England. A broad belt including most of the tracks of the selected depressions receives an average of 10 to 15 mm. Only in S.E. England does the average rainfall fall below 10 mm, there reaching a minimum of about 7 mm. near the Thames estuary. To the north of the track of the depressions rainfall falls off rapidly and the uniform values of 4 or 5 mm. found over North Scotland can be attributed to showers in the polar air north of the centre. As would be expected the two stations in the Welsh hills receive considerably more rain than the places on the coast and in England. On the other hand, at Eskdalemuir rainfall is no greater than at corresponding places on the east and west coast; the relatively light winds north of some centres would be unfavourable for significant orographic rain, and the easterlies to the north of other centres may have no very marked orographic rainfall increment at Eskdalemuir which lies west of the watershed.

In order to form a clearer picture of the rainfall in relation to the track of the depression an attempt was made to estimate the average rainfall over areas centred 50, 150 and 250 miles to the north and south of the track of each centre. Western and eastern areas of the United Kingdom were treated separately and the resulting rainfall profiles are shown in Fig. 2. The curves for eastern and western districts are very similar. The maximum rainfall is slightly north of the depression track in eastern districts and slightly south of the track in western districts; this displacement towards the strongest on-shore winds has an obvious orographic explanation. The same effect is apparent in Fig. 1 when the east coast rainfall has a maximum in the north near Middlesborough whereas the west coast maximum is farther south over South Wales. It is clear from Fig. 2 that in the absence of orographic effects the average rainfall has a maximum very close to the depression track and falls off symmetrically at least to a distance of 200 miles where the amount is about half of the central maximum.

Fig. 3 shows the percentage of the 45 depressions which gave less than 0·1 mm. of rain at stations in the British Isles. The main features are:-

(a) At stations in England measurable rain accompanied nearly all of the depressions; over a belt across the Midlands and Southern England rain fell from every depression examined.

(b) Even in Scotland which frequently lay outside the depression's circulation there was measurable rainfall during the passage of most of the depressions. This can be attributed to showers in the cold air mass north of the centre.

(c) There is no difference in the number of cases without measurable rain between stations in mountainous areas and elsewhere.

Fig. 4 shows the standard deviation of the rainfall of individual depressions from the mean in Fig. 1; histograms are also included for selected stations. The fairly wide range of variation in rainfall amounts is shown. The standard deviation tends to increase with the amount of rain but over central England close to the paths of most of the depressions the variability is a smaller fraction of the average rainfall than elsewhere.

4. Characteristics of individual rainfall distributions

The individual rainfall charts showed a wide variety of patterns. About half of the depressions gave a definite band of heavier rain along their track and in some cases this was exceedingly well marked. (Fig. 5 illustrates this for the depression of 6-7 April, 1949).

/In

In other cases the rainfall pattern was irregular (see Fig. 6 - depression of 7 January, 1948). Some of the irregularities were due to orographic effects and some to the occurrence of local thunderstorms, but others must have been connected with irregularities in the structure of the depression. (No reports from mountainous areas were employed in constructing the isohyets in Figs. 5 and 6 so that the orographic maxima over high ground are not adequately shown). Orographic maxima in northeast England and South Wales, as illustrated in Fig. 6, occurred in a number of cases. The east coast maximum is aided by rainfall from the occlusion which often becomes slow-moving over the area as the depression passes further south. One case included in the present series is that of the depression of 11 August, 1948 which gave rise to the Tweed Valley floods; in this case the depression became slow-moving off the east coast and rainfall was unusually prolonged.

A study was made of the surface pressure and thickness charts with a view to ascertaining the factors which determine whether or not a definite rainfall band should occur along the track of the depression. No clear-cut relationships were found. There was a recognisable tendency for the depressions moving faster than 30 kts. to give a definite band of heavier precipitation (15 out of 19 such cases gave a well defined band) and for depressions moving slower than 25 kts. to give an irregular rainfall pattern (only 3 out of 16 such cases gave a definite rainfall band). The clear-cut rainfall band is also more common in winter and spring than in summer and autumn; it accompanied 70% of the cases from November to April but only 31% of the cases from May to October. The thickness pattern gave no indication good enough for use in forecasting, although there is a tendency for the depressions which give the definite bands of heavy rain to be associated with straight thickness lines more often than those which gave irregular rainfall distributions. The thermal gradients also tend to be slightly stronger with the former.

5. Variation in rainfall amounts in relation to synoptic parameters

A brief study was made of the rainfall of individual depressions in relation to selected parameters characteristic of the synoptic situation. A similar study of the rainfall of depressions passing eastward between Scotland and the Faeroes had given results of little pratical value in forecasting and it was therefore considered desirable to confine the present investigation to the simplest and most direct relationships.

As a measure of the general rainfall from each depression the average of the rainfall at 12 stations spread over England and Wales was calculated. This average had a correlation of about +0·8 with a similar average for 4 stations in southeast England and it can be taken as reasonably representative of the "raininess" of a particular depression.

Relationships were investigated between the rainfall parameter, R, and the speed of the depression, the surface pressure field and the thickness pattern.

The rainfall was found to have a negative correlation of -0·30 with the speed of the depression and a negative correlation of -0·46 with its central pressure (the correlation with the vorticity based on the surface pressure field was +0·45). It was also found that the "wet" depressions tended to have a component of motion towards the north (i.e. to move from S.W. to N.E.) whereas those moving towards E.S.E. or S.E. were "drier". The correlation between the rainfall and the direction of motion is -0·43.

The 1000-500 mb thickness pattern also provides some significant relationships with the rainfall. The rainfall has a correlation with the orientation of thickness lines corresponding to the correlation with the direction of motion and of about the same magnitude; there is also a correlation of about +0.3 with the thickness gradient over the centre at right-angles to the direction of motion. As these relationships are no closer than those between rainfall and the simpler parameters taken from the surface chart, they have little practical importance. There is however some theoretical interest in Fig 7 which shows the mean 1000-

/500 mb.

500 mb thickness charts for the 10 most rainy and 10 least rainy depressions. It is interesting to note that the average of the rainy cases shows a definite thermal trough just west of the British Isles whereas the average of the least rainy cases shows a straight thermal gradient presumably to the west of a trough-line over Europe.

The partial regression equation has been calculated between the rainfall R and the three parameters

(a) V - the speed of the depression (kt.)

(b) p - the central pressure of the depression (mb)

(c) θ - the orientation of the track of the depression in degrees from north.

The regression equation is:-

$$R = 15{\cdot}9 - 0{\cdot}18\,V - 0{\cdot}23\,(p - 1000) - 0{\cdot}092\,(\theta - 90) \qquad (1)$$

The total correlation between R and the result of the regression equation is 0·60. Such a correlation is not high enough to be of much direct practical value in forecasting, particularly when the parameters themselves have to be forecast. However, the general relationships of rainfall with the speed, direction and depth of the depression are useful as "background knowledge" to the forecaster.

The weakness of an equation such as (1) for forecasting rainfall is illustrated by comparing the actual values of the average rainfall R with rainfall calculated from equation (1) assuming the parameters V , p and θ are known. If the rainfall amounts are divided into convenient groups 0, Tr to 1 mm, 1·1-5 mm, 5·1 to 10 mm, 10·1 to 20 mm, 20·1 to 30 mm and greater than 30 mm the contingency table is as given below.

Table I. Comparison between average rainfall and value given by regression equation (1)

Rainfall given by regression equation (1)	Average rainfall R: 0	Tr to 1 mm.	1·1 to 5 mm.	5·1 to 10 mm.	10·1 to 20 mm.	20·1 to 30 mm.	More than 30 mm.
More than 30 mm.	0	0	0	0	0	0	0
20·1 to 30 mm.	0	0	0	0	0	1	0
10·1 to 20 mm.	0	0	0	9	21	2	2
5·1 to 10 mm.	0	0	2	3	1	-	-
1·1 to 5 mm.	0	0	2	2	0	0	0
Tr to 1 mm.	0	0	0	0	0	0	0
0	0	0	0	0	0	0	0

/The

The regression equation gives 27 values correct, 16 in error by one class and two in error by two classes. On the other hand if it were predicted that the precipitation would always be between 10 and 20 mm, there would be 22 corect values, 17 in error by one class and 6 in error by two classes. This small improvement is achieved when the parameters are assumed known and the regression formula is tested on the data used to derive it. The improvement from use of the regression equation would be even less in practice.

6. Conclusions

Depressions moving eastward across England north of the Bristol Channel and Thames estuary give significant rainfall at all stations in England and Wales on nearly all occasions. The average rainfall along the track of the depression is about 15 mm. and falls off to about half this value at a distance of 200 miles. Rainfall amounts are much the same on low ground in the east and west of Great Britain but the heaviest falls are slightly further north on the east coast than in the west.

A well defined band of heavier rainfall is found along the track of the depression with about half of the depressions. It is most common in winter and spring and with fast moving systems.

The investigation confirms the recognised tendency for rainfall to be greater with the deeper and more intense depressions, with those that move more slowly and with those which have a component of motion towards the north. It is not of practical forecasting value to use any statistical formulae to express these facts.

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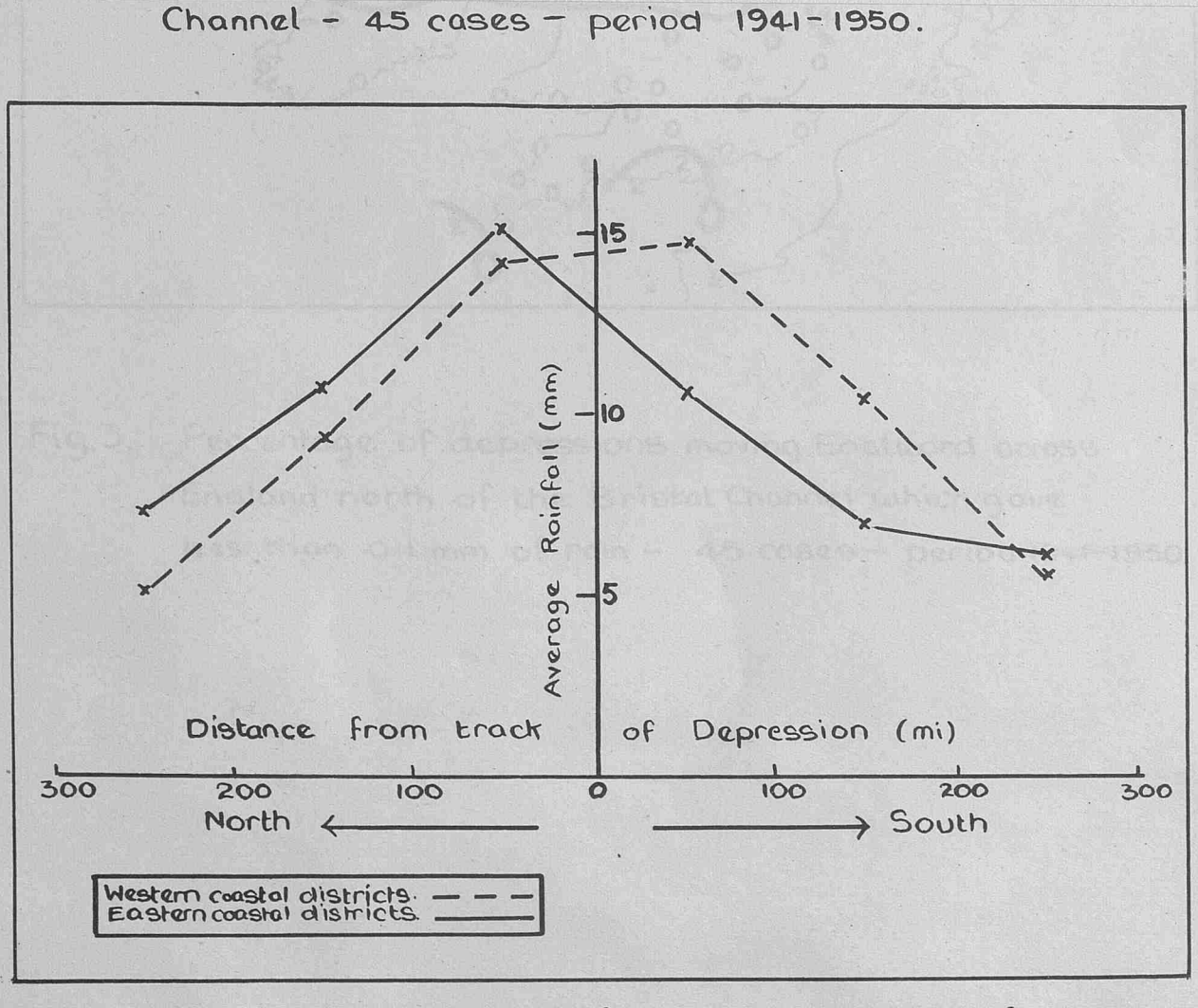


Fig.2. Average profile of rainfall across the track of a depression crossing England.

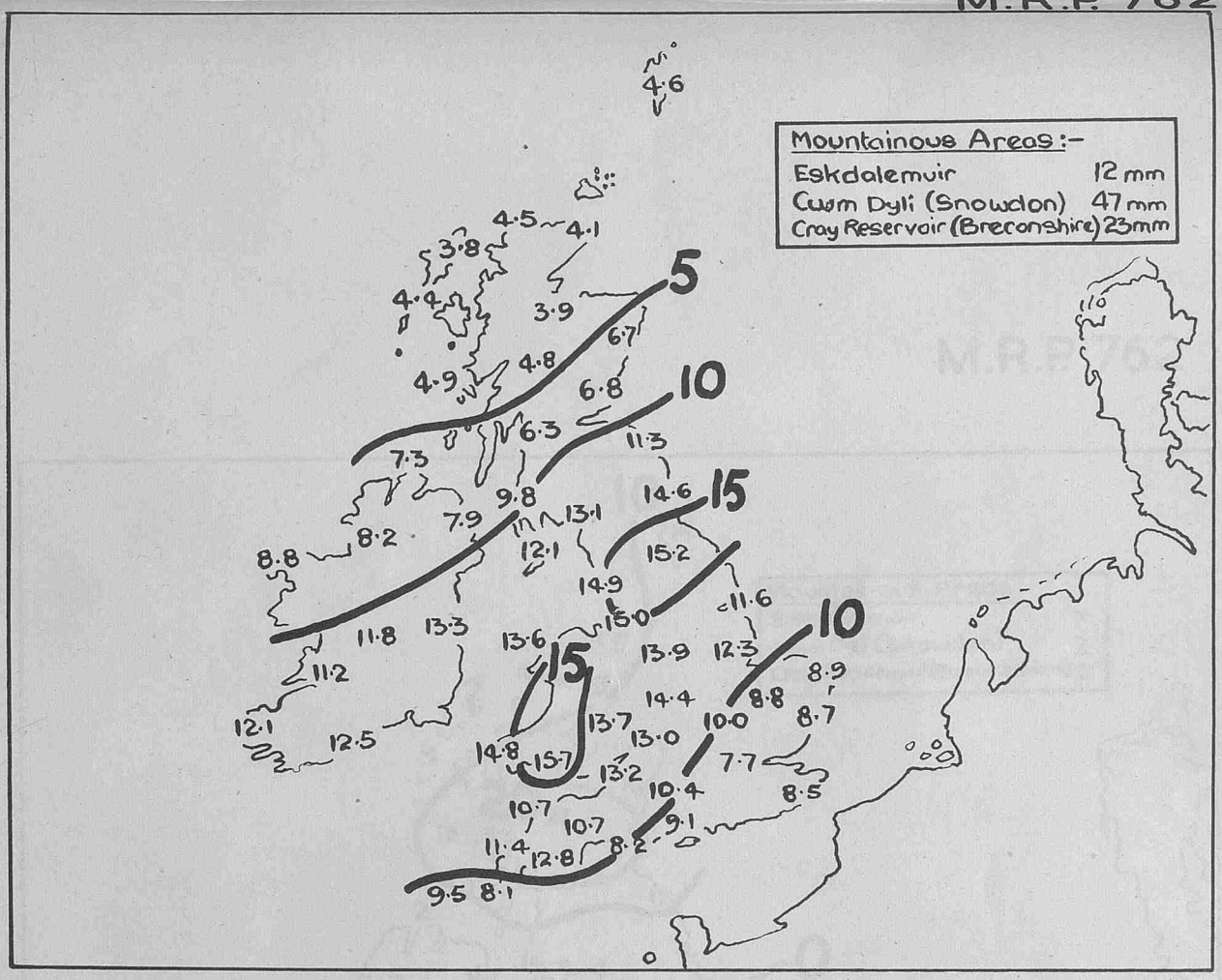


Fig. 1. Average rainfall (mm) from depressions which moved eastward across England north of the Bristol Channel - 45 cases - period 1941-1950.

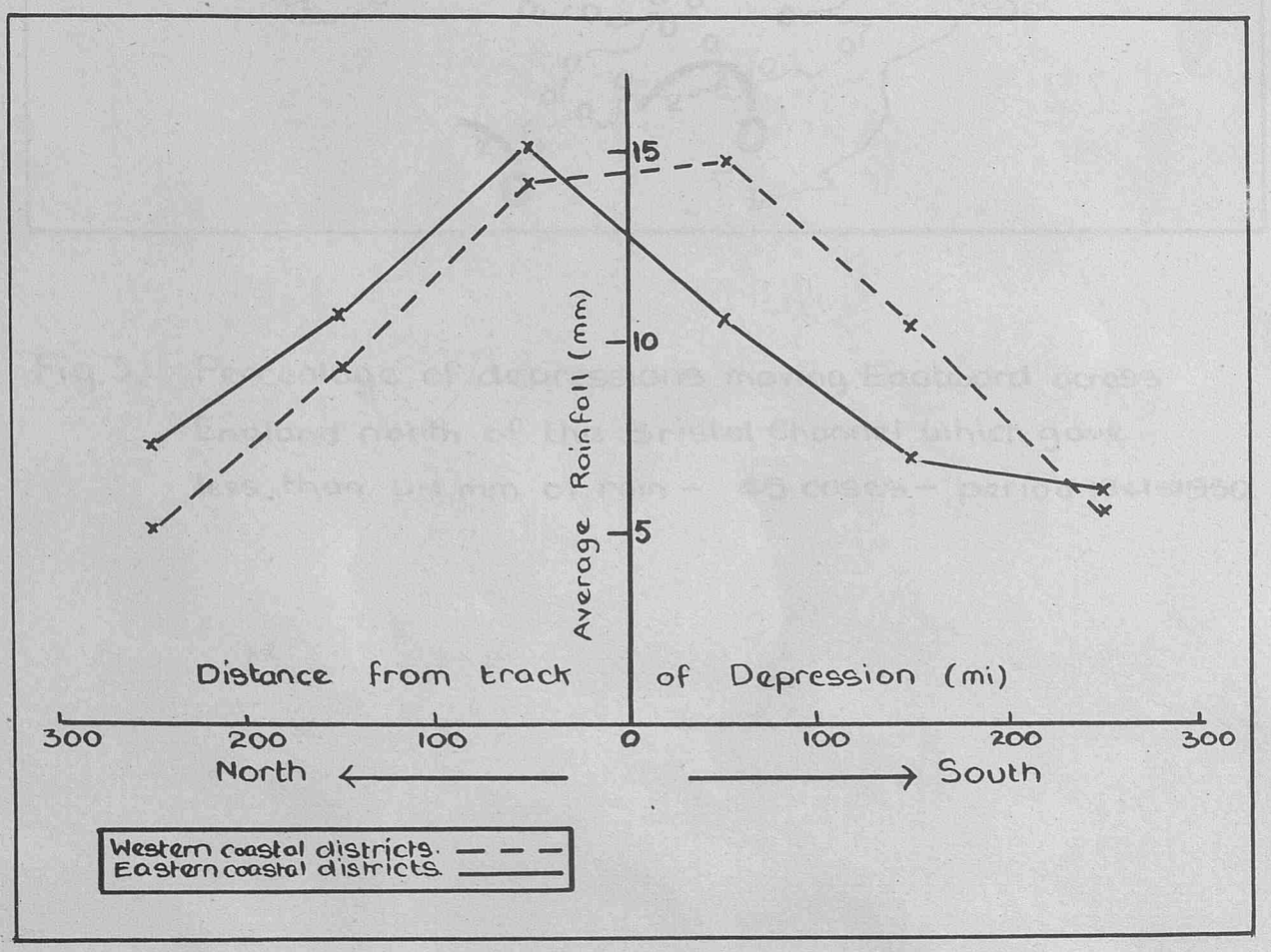


Fig. 2. Average profile of rainfall across the track of a depression crossing England.

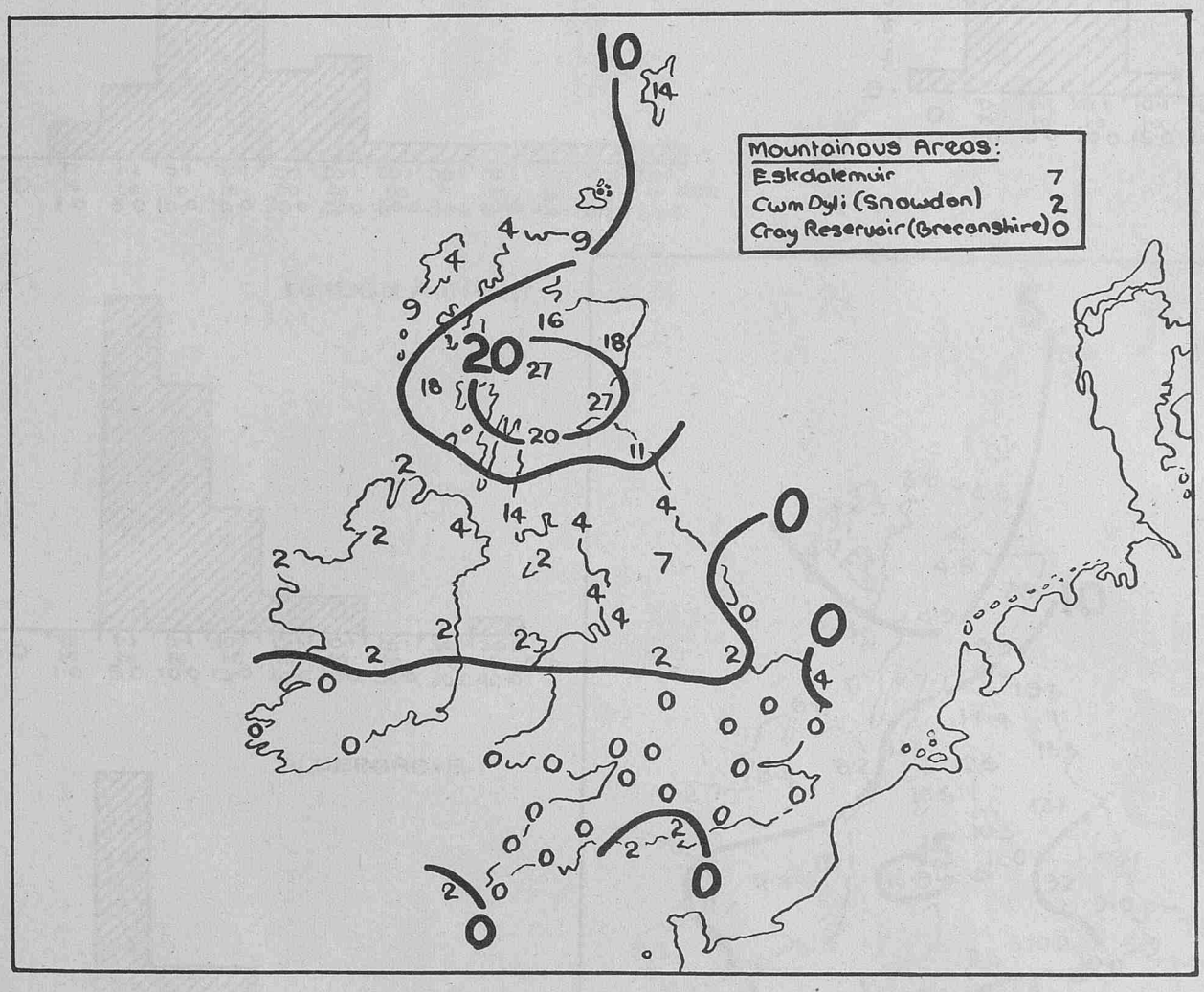
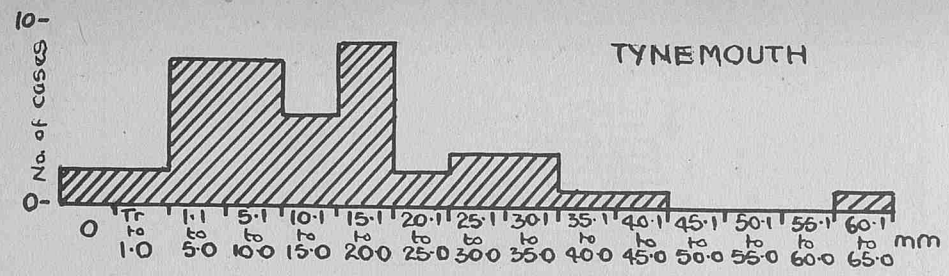
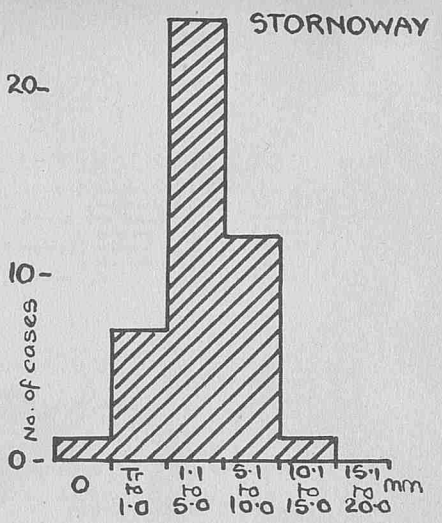
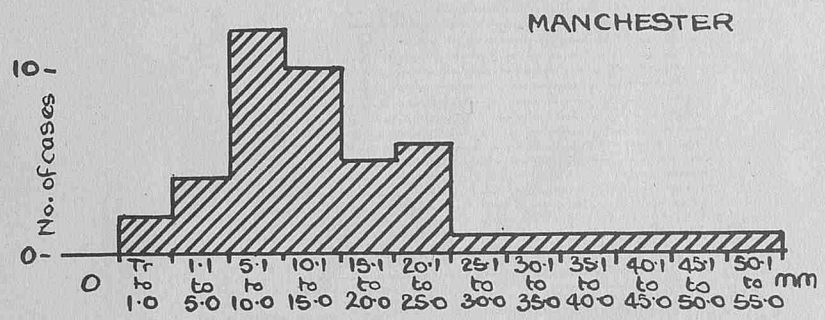
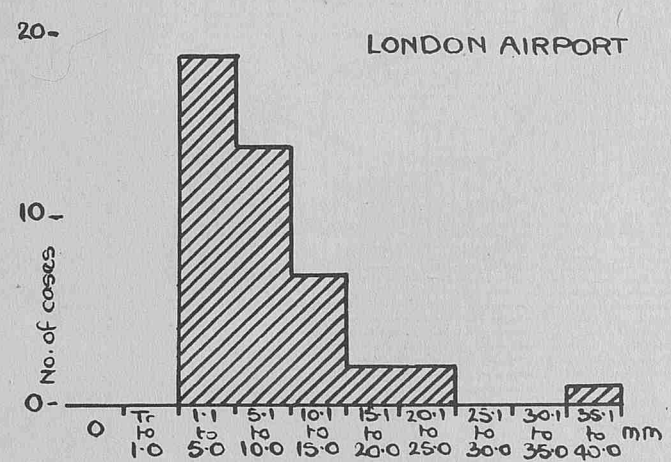


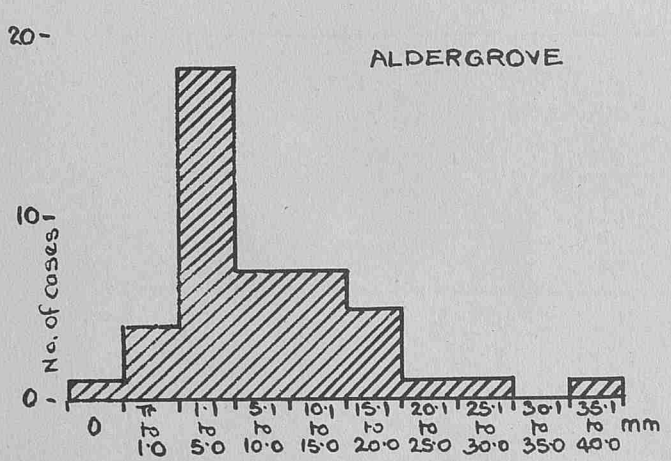
Fig.3. Percentage of depressions moving Eastward across England north of the Bristol Channel which gave less than 0·1 mm of rain – 45 cases – period 1941-1950.

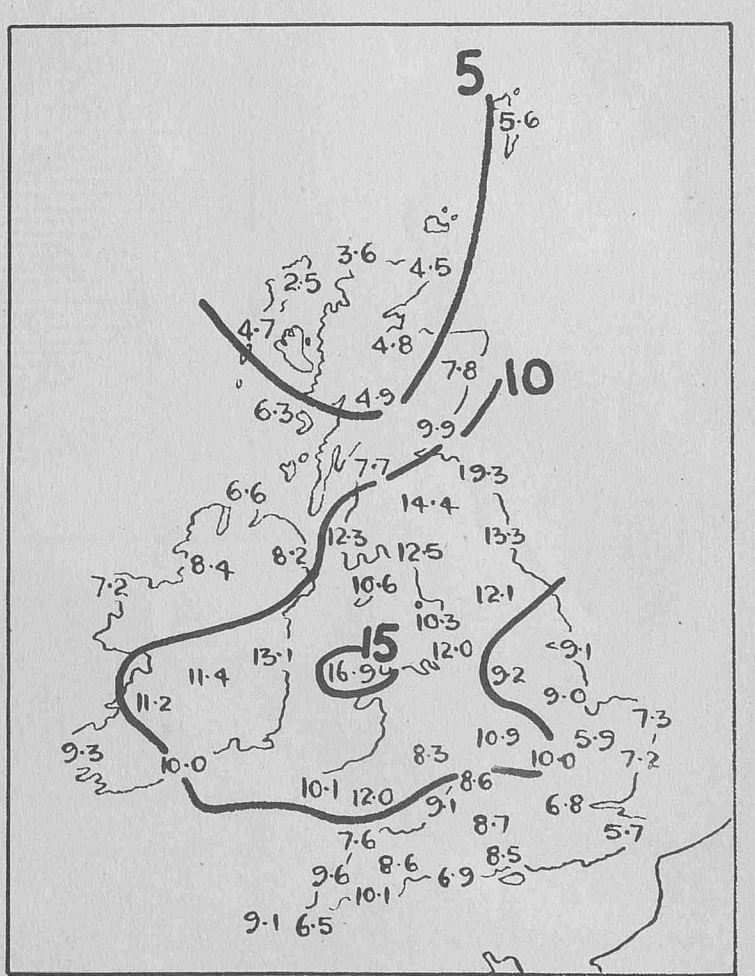


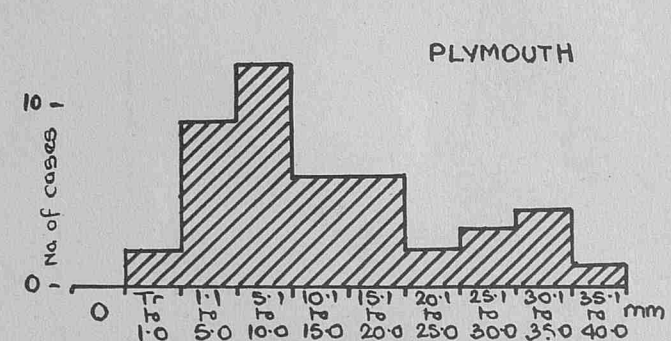


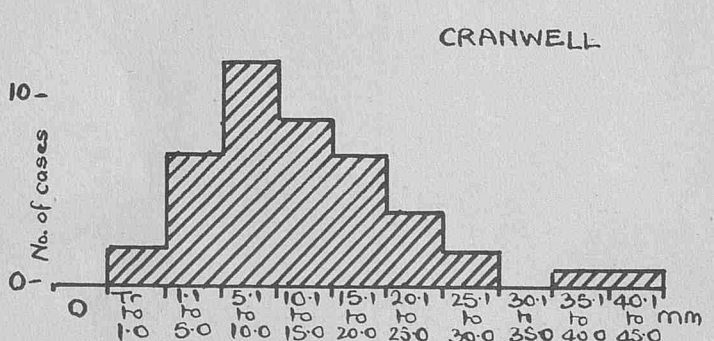






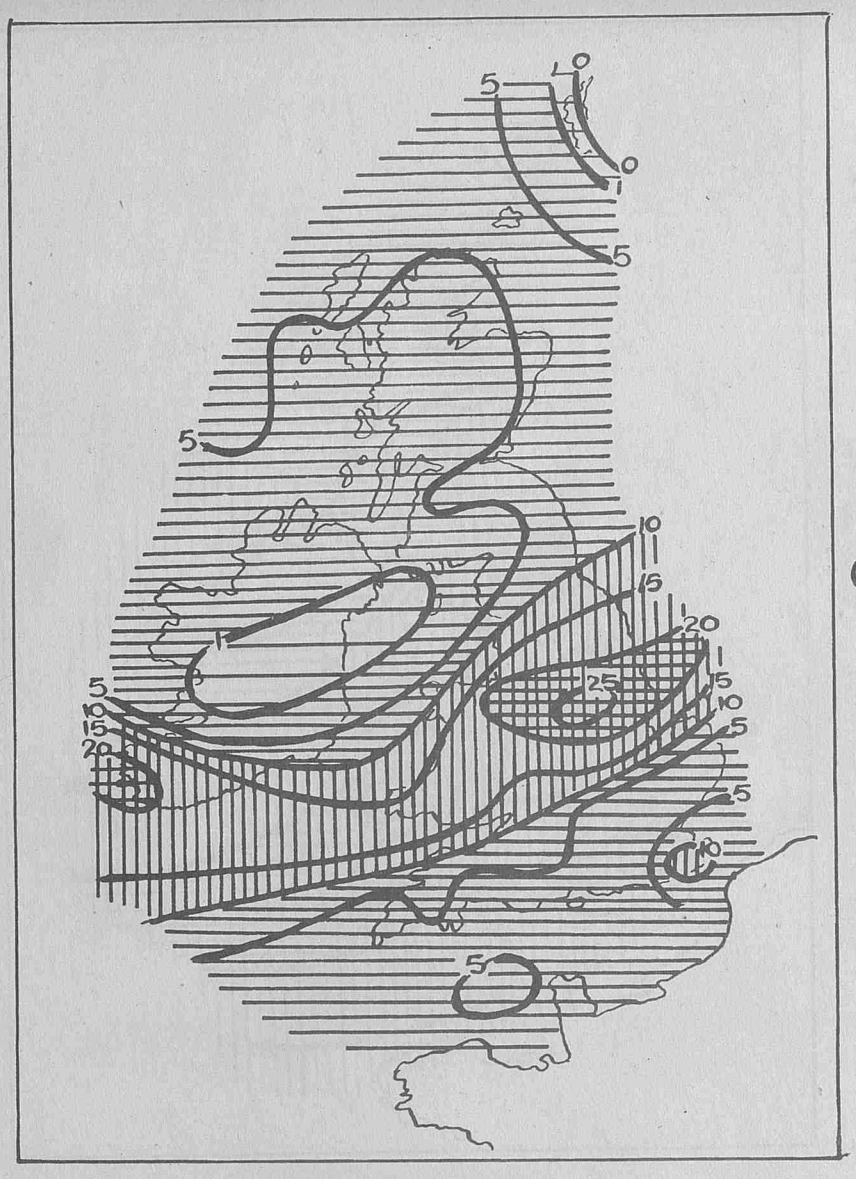






Depressions which crossed England north of the Bristol Channel.
45 cases – period 1941–1950.

Fig.4. Standard deviation of rainfall of individual depressions (mm) and frequency diagrams of rainfall totals from individual depressions.



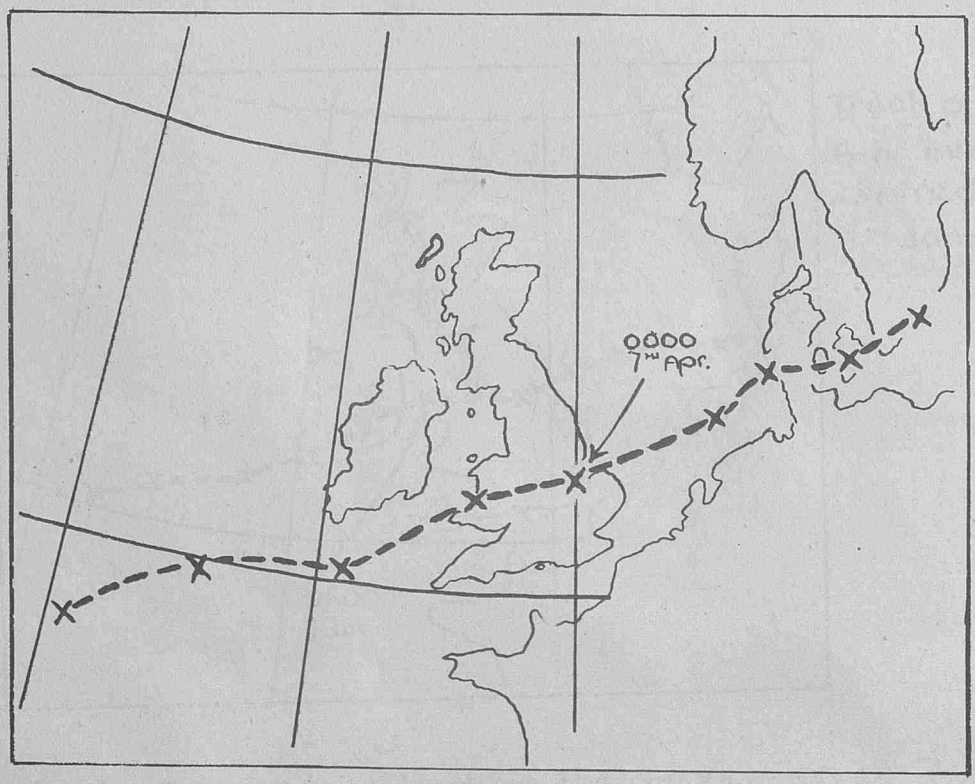
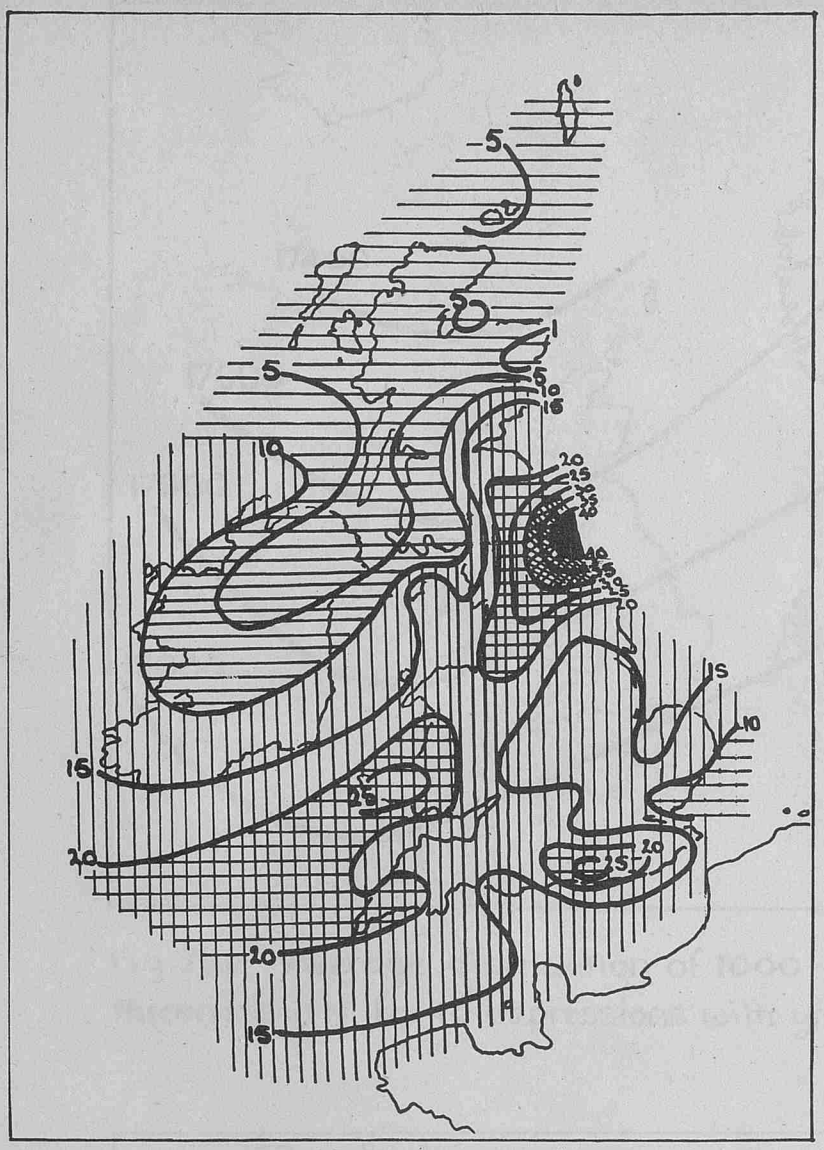
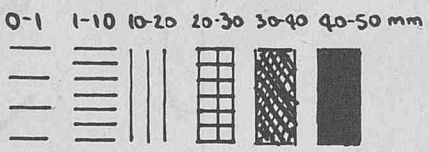
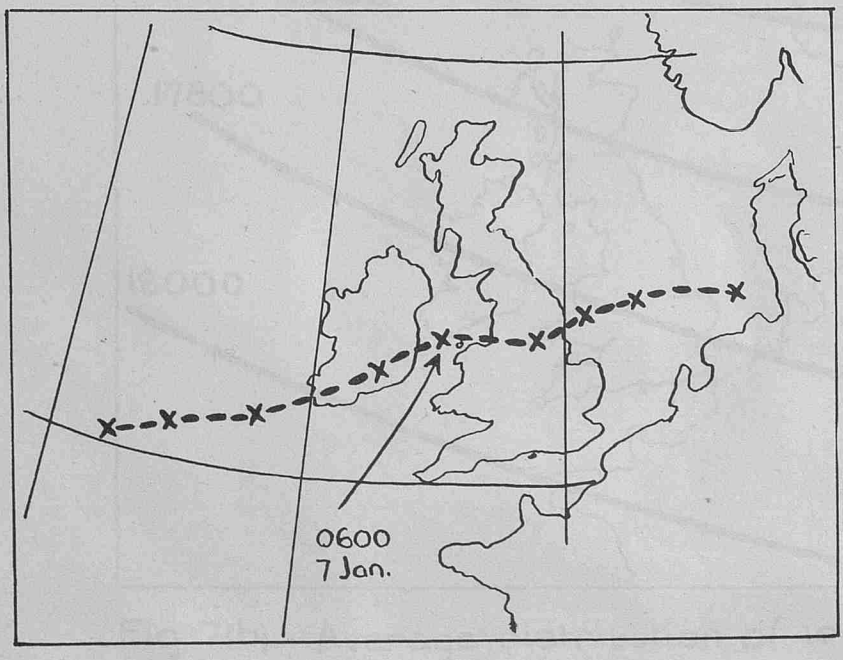


Fig. 5 Rainfall and track of the depression of 6-7 April 1949





Rainfall:
period- 21hr 5 Jan. 1948
to 09hr 8 Jan 1948
(based on telegraphic reports.)



Track of depression
6-hr intervals
centred 0600hr
7th January 1948.

Fig 6. Rainfall and track of the depression of 7 January 1948.

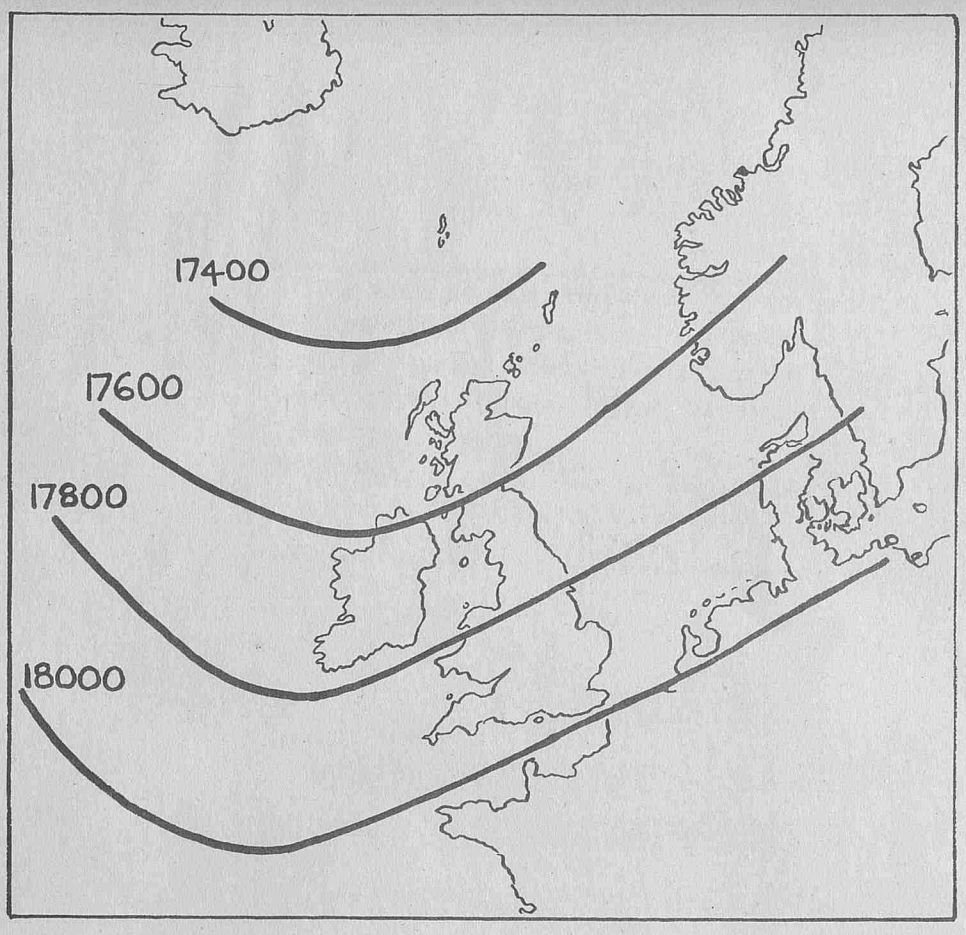


Fig. 7(a). Average distribution of 1000-500 mb thickness (ft) for 10 depressions with greatest rainfall.

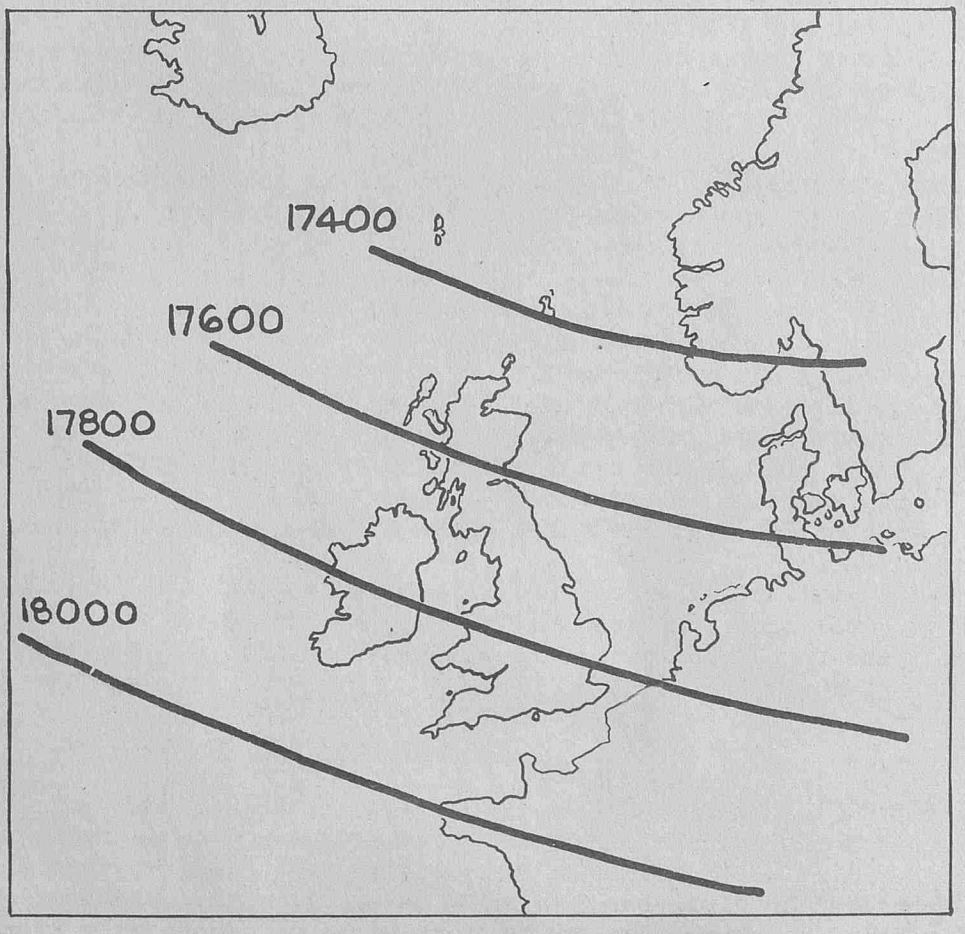


Fig. 7(b). Average distribution of 1000-500 mb thickness (ft) for 10 depressions with least rainfall.

11-60
900Kms

M.R.P. 764
S.C.II/125
31st October, 1952.

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A copy is available in the Library of the Meteorological Office."

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

COLD POOLS: A STATISTICAL AND SYNOPTIC STUDY

by
E.J. Sumner (Meteorological Office)

1. Introduction

1.1. The cold pool is one of the more common synoptic models which have come to be recognised since the introduction of upper-air charts into forecasting practice. There is already a considerable amount of literature on the subject, most of it however being concerned with studies of the life-history and detailed structure of individual pools. Little work of a statistical and synoptic nature appears to have been done, at least in recent years (before which the observational coverage was scarcely adequate), and the present study was undertaken in order to fill in some of the gaps.

1.2. A cold pool may be defined as a mass of cold air in depth entirely surrounded by relatively warm air, and shows up as one or more closed lines in the thickness isopleths for any fairly deep atmospheric layer. The area within the outermost of these closed lines may for present convenience be identified with the cold pool. For statistical purposes some limitation as to size is necessary, otherwise the usual zonal distribution of thickness lines associated with the entire "circumpolar vortex" may itself be classified as a cold pool. At the other extreme a minute and ephemeral pocket of cold air contained within one closed thickness line (the presence of which may depend on the interval at which the lines are usually drawn, and may in fact be drawn in arbitrarily owing to observational shortage) is of slight interest. A similar definition to the above was given by Douglas (1947).

Various related terms are in use in the literature; these include cold domes, cold poles, cold drops (Kaltluftropfen), cold lows, cold highs and cold "cut-offs". A cold pool, in the sense used here, may occur with any of these phenomena.

2. The data and the measurements

2.1. The charts used in this study were the 0300 and 1500Z circumpolar 1000-500 mb. thickness maps prepared daily in the Meteorological Office, Dunstable. The data are for the 5 year period from September, 1946 to August, 1951, and for the area between longitude 60°W to 30°E and south of latitude 80°N. These limits were chosen so as to exclude the semi-permanent winter cold pools

/over

over the Asian and American continents, and to confine the investigation to those features which might be of direct concern to the British Isles and its environs.

2.2. Thickness lines on the circumpolar charts are drawn at intervals of 200 feet (the even hundreds). Only well-defined and fairly persistent pools were considered, the minimum requirements being that there should be two or more closed thickness lines surrounding the pool which should appear on at least two successive 0300Z charts, the closed lines lying entirely within the area delineated above. This criterion enables one to avoid the extremes mentioned in the introduction, in fact it is perhaps a little too restrictive in that some pools contained within a single closed thickness line are certainly of more than passing importance in the synoptic picture. Actually, within a particular spell, one-day interruptions during which the pool was represented by a single closed line at 0300Z were included (5 cases in all), thus allowing for a period of temporary waning. (In a few cases an intense pool was represented by a single thickness line, or even degenerated to a trough, on the intermediate 1500Z charts. These were, of course, allowed on our criterion.)

The central position of each pool was recorded to the nearest degree of latitude and longitude, and also the interpolated value of the 1000-500 mb. thickness at the centre, to the nearest 50 feet. The "centre" was taken as the approximate centre of gravity of the area of the pool (no matter how irregular in shape) and was estimated by eye. The intensity of the pool was also recorded on the scale: one closed thickness line, intensity one; two closed lines, intensity two, ... and so on.

3. Some general statistics

3.1. Within the 5-year period under consideration the total number of spells was 75, ranging from 2 to 10 days' duration, the average being about 3·0 days. A spell is said to be of n days' duration if the same cold pool (beginning and ending with intensity two or more, but possibly with one day interruptions of intensity one) appeared on n successive 0300Z charts. The total number of individual pools involved (i.e. occurrences on 0300Z charts) was 224. By definition the minimum intensity was generally two, with a few of intensity one as explained above; the maximum intensity was five.

3.2. The positions and central thicknesses of all the pools were marked on a map, the seasons being distinguished (Figs. 1 and 2). The greatest concentration is over Europe at all seasons, but in spring and summer there are several other clusters, more notably in the western Atlantic and in the area between NE. Greenland and N. Scandinavia. The almost complete absence* of pools of intensity two or more just west of the British Isles, around and to the east of Iceland, and in the western Atlantic (summer and autumn only) is particularly striking. These gaps possibly arise in part from the shelter provided by the (warm) American continent in summer and the Greenland ice-cap at all seasons. However, the preference of blocking (usually warm) highs for the north-eastern Atlantic in May and June is also associated with the paucity of pools in the Iceland region (see section 4.1, later).

The movement of each pool is shown by a connecting line and an arrow; individual spells are identified by their commencing date. Most of the pools were fairly slow-moving (usually less than 500 miles/day), and any rapid displacements were seldom continued beyond a day. In particular the pools in high latitudes showed no tendency to come southwards; these latter are presumably the cold "poles" of the Northern Hemisphere and in what follows they were separated from the others (and later left out of account), a convenient line of separation being the 65°N-parallel.

/3.3.

* Actually a very small number of pools of intensity two did occur in these areas but did not last beyond a day.

3.3. Figure 3 shows the frequency distribution of pools by months (5 Januaries, 5 Februaries, etc.), pools north and south of 65°N being distinguished. The number of spells by months is also indicated (spells overlapping into the adjacent month were classified in the month containing the larger part of the spell, or in the earlier month if it were equally divided between them). It is evident that, within the area considered, these intense cold pools are largely spring and early summer phenomena, an outstanding maximum occurring in May and June.

3.4. The frequency of spells in days (see 3.1 for definition of a spell) is given in Fig. 4. A persistence of two days, the minimum required by our definition, is very much more likely than any other. (There were actually very few cases in which a pool of intensity two appeared on only a single chart without previous or succeeding trace on an 0300Z chart, so that the maximum frequency of duration of major pools was undoubtedly two days.) This applied to all seasons. The average spells north and south of 65°N were 3·9 days and 2·8 days, respectively. North of 65°N there were 51 pools (13 spells) in all, the corresponding figures south of this parallel being 173 pools (62 spells).

3.5. The frequency of intensity is shown in Table 1.

Table 1

Position of pool	Intensity of pool			
	2	3	4	5
South of 65°N	137	31		
North of 65°N	34	10	5	2
Total	171	41	5	2

There were relatively few cases of intensity more than two, and all those of intensity greater than three were north of 65°N. There was no relationship between the initial intensity and the subsequent duration of a spell, although there was a small positive correlation between the duration and the average intensity within a spell.

POOLS SOUTH OF 65°N.

We shall only be concerned in the remainder of the paper with those pools south of 65°N, partly for the reason given previously and also because a more detailed study of weather, cloud, etc. attempted here is not possible in higher latitudes owing to observational shortage.

4. Mode of formation and disappearance

A scrutiny of all the thickness and surface charts was made in order to study in synoptic terms how each pool formed and disappeared, with the following results.

4.1. The greatest number of pools starting a spell - 49 out of 62 - were formed as a result of a partial or complete cutting-off of the cold air near the southernmost extremity of a cold trough. The cold trough was usually fairly slow-moving and of large amplitude at the time of cutting-off or was increasing in amplitude, the low-latitude part slowing down still further or

/stagnating

stagnating while the high-latitude part moved on. A good example of this cut-off process is given in Figs. 5a, b and c. In this example there was marked anticyclonic building across the middle of the trough with a cyclone maintained to the south in association with the developing pool. A certain amount of warm advection from the west round the top of the anticyclone completed the cutting-off. There were, however, many variants of this basic model mainly depending on the degree of development of these surface features. In eight of the 49 cases the depression predominated, and warm advection from the east round the northern flank of the low seemed to be responsible for cutting off the pool; on 19 occasions the low was vestigial and marked anti-cyclonic building and presumably subsidence across the neck of the trough appeared to be the main agency; in a further eight cases both these effects seemed to be important. In two cases the cold pool actually formed "over" the surface high but was then quickly transferred - almost "jumped" - south to come more in association with the low. The remaining 14 cases were due to a more subtle combination of advectional and dynamical thermal processes (including local cooling at the centre of the pool and warm-air advection from the west round a depression in high latitudes) operating in different parts of the field. It is not suggested however that the above remarks adequately represent what goes on, or that such thermodynamical processes are primarily responsible for this type of development.

Of the remaining 13 pools, six were advected into the area from outside (five of them into the western Atlantic from the NW., one from the east into the Baltic); and seven were formed in association with a surface low, partly by a sort of cutting-off in rather higher latitudes than usual but with a certain amount of cooling in situ as an important contribution (these cases are ringed in Figs. 1 and 2). As will be seen later these latter occurred almost entirely in summer when synoptic patterns were shifted north, and the difference between these and the more typical cut-off low may not be fundamental. However, these cold lows were not associated with a "blocking" of the westerlies, whereas the "cut-off" pools almost invariably were. In this latter connection, the indirect agreement with the results of Brezowsky et alia (Tellus, 1951) is striking. These authors found a pronounced maximum of blocking highs with axes in the sector 20°W - 10°E, to occur in May and June (70 years' data).

4.2. The greatest number of pools (33 out of the 62) disappeared (or were reduced in intensity and therefore no longer considered) by warming more or less in situ. Almost exactly half of these disappeared from the charts within 24 hours; the rest lingered on at intensity one for several days before finally degenerating. (In one case the original pool regenerated again several days later and was considered as another spell; in this instance the mode of formation was classified as that of the original.)

In nine further cases the pool moved so as to be absorbed into the colder air of higher latitudes, two were reabsorbed into the original cold trough by renewed advection from the north, and a further two seemed to be reabsorbed in this way by local cooling to the north of the pool. The remaining 16 disappeared as a result of a combination of these factors, warming in situ being one important agency in most of them. In all these classes too, about half of the pools remained at intensity one for a further day or more, before finally disappearing from the charts.

5. The surface-pressure systems associated with cold pools

5.1. Each pool was classified according to the surface pressure field in its locality. The following six surface types were classified as associated, viz., a low (L), a trough (T), a high (H), a ridge (R), a slack area or col (C)* and a fairly straight run of isobars more or less mid-way between a large high and a large low (S). Three of these types are usually called the cold low (L - type) the cold high (H-type) and the cold drop (S-type), respectively, but the terms cold trough or cold ridge are not used in this context. Although in

/general

* Actually only one case of a col was recorded, the rest were in a very weak and irregular pressure field.

general little difficulty was found in classifying, a few cases were borderline and a certain amount of arbitrariness is inevitable. The results of this classification on a seasonal and "land-sea" basis are given in Table 2.

These results include all the pools south of 65°N. However, those formed in association with a surface low without having been cut off at the bottom of a cold trough, although included in the table, are also distinguished from the rest by their number being put in brackets after the total in each category. These cases occurred almost entirely in summer over the sea; they are not all classified as L-types because in a few cases the pool having formed in association with a surface low subsequently became associated with a different surface-pressure field.

From the totals in each column it is evident that in winter and autumn cold pools are more frequent over the land than over the sea, and vice versa in spring and summer. However, if the cold-low type referred to above (figures in brackets) are excluded, even in summer cold cut-off pools are more frequent over land than sea.

By far the greatest number of cold pools were associated with surface lows at all seasons, the second greatest being "cold drops". Associated troughs and slack areas were next in almost equal proportions; ridges and anticyclones were in a minority. Lows and troughs were relatively more frequent over the sea, and all other types (highs and ridges especially) over land. Except in the case of highs and ridges, the seasonal distribution was in keeping with Fig. 3, i.e. with a maximum in spring and summer.

5.2. There were 33 spells in which the same type of associated surface-pressure system predominated throughout (i.e. in which spells of three days or more duration were allowed to have at most one day of a different type), 25 of these being pure types (i.e. no interruptions permitted). The individual figures (pure types in brackets) were: L-type 18 (16), T-type 4 (3), H-type 1 (1), R-type 2 (1), C-type 2 (1) and S-type 6 (3). A further 16 spells had the ingredients L, T and S predominantly (almost invariably two at a time), and 5 were combinations of R, H and C. The remaining 8 (of the 62 total) consisted of other combinations, two or three at a time, of all the types.

Within a spell extreme changes from an associated high or ridge to a low or trough, or vice versa, were very infrequent as the above figures imply. When they did occur pressure gradients were usually slack (borderline C-type) with a high general level of pressure; in such circumstances comparatively small pressure changes would suffice to produce definite ridges and troughs (or even weak lows) on alternate days.

There was a tendency for the straight isobars of the S-type to become more cyclonically curved with time, with a change to a T- or even an L-type; in fact, over the sea, there were no pure S-types - all changed, mostly to L- or T- types.

5.3. For each class of pool the seasonal and land-sea differences between the average durations of the spells were generally small; a few large differences were not considered to be significant owing to the small number of cases involved. However, for all the data, there was some suggestion of a seasonal variation in the length of a spell, for in winter the average duration was 3·9 days (8 spells), spring 2·5 days (24 spells), summer 2·8 days (20 spells) and autumn 3·4 days (8 spells). The average duration was slightly longer over the land (2·9 days, 29 spells) than over the sea (2·7 days, 33 spells), but these differences are so slight that it is doubtful whether any significance can be attached to them.

/Table 2.

Table 2.

Associated surface-pressure system	Winter (Dec. - Feb.)		Spring (March - May)		Summer (June - August)		Autumn (Sept. - Nov.)		Total for Year		
	Land	Sea	Land	Sea	Land	Sea	Land	Sea	Land	Sea	Both
L	9	2	7	23 (1)	4	18 (14)	7	1	27	44 (15)	71 (15)
T	0	2	3	4	7	6	1	1	11	13	24
H	1	1	0	0	1	0	3	2	5	3	8
R	3	0	1	0	5 (1)	0	3	1	12 (1)	1	13 (1)
C	4	2	3	2 (1)	5	4	2	1	14	10 (1)	24 (1)
S	5	2	11	4	3	6 (1)	2	0	21	12 (1)	33 (1)
Total	22	9	25	33 (2)	25 (1)	34 (15)	18	6	90 (1)	83 (17)	173 (18)

/5.4

5.4. Table 3 shows the frequency distribution of distances between the centres of the pools and those of the associated surface-pressure systems.

Table 3.

Surface-pressure system		Distance between centre of pool and centre of pressure system: hundreds of miles										Totals
		0	1	2	3	4	5	6	7	8	9	
LOWS	(Land	1	5	5	8	4	2		1		1	27
	(Sea	6	13	15	6		2	1	1			44
	(All	7	18	20	14	4	4	1	2		1	71
HIGHS	(Land			2	1		2					5
	(Sea								3			3
	(All			2	1		2		3			8

Most of the lows were 300 miles or less from the cold pools, although few of them (especially over the land) were quite concentric with the pool. The average distance between the centre of the pool and that of the associated low was about 300 miles over the land and 220 miles over the sea. All lows more than 300 miles away were situated on the sector between SE. and NE. from the associated pools. The remainder were almost equally distributed in direction with respect of the centre of the pool.

The corresponding averages for the few cold anticyclones were 340 miles over the land and 700 miles over the sea. The three latter were all west of the associated pools: the former on the other hand were well scattered with respect to the centre of the pool.

Cold drops (S-type) were on an average closer to the associated low (average departure 600 miles) than to the high (average departure 940 miles). This difference is presumably connected with the average sizes of pressure systems, lows being smaller than highs as a rule. Both average distances were greater over the land than over the sea by about 20%.

5.5. The surface pressure was measured, to the nearest millibar, at the centre of the pool (P) and at the centre of the associated high or low (p). These results are set out in Tables 4a and 4b.

As one would expect, the averages of actual pressure and pressure anomaly for each type of associated surface-pressure system (Table 4a) are in descending order of magnitude from type H, to R, C and S (almost equal), T and finally L. For all the data, the average surface-pressure both at the centre of the pool and at the centre of the associated surface system, was lower over the sea than over the land: the corresponding average pressure anomaly changed from positive to negative or became more negative from land to sea. This was not however true for each of the individual classes, the C-type in particular showing opposite characteristics both in respect of pressure and pressure anomalies.

The averages for all the data are of course heavily weighted by the preponderance of the L-type, especially over the sea (Table 2) where, also, the diminution of mean surface pressure and the increased (negative) anomaly is especially marked. Thus associated cyclonic circulations tend to be more intense and relatively more frequent over the sea. A few cases are on record where a cold pool associated with more or less straight surface isobars

/or

or with a weak surface low, moved from the land to the relatively warm sea in winter. In each case there was a noticeable increase in the associated cyclonic circulation at the surface (with the formation of a low if one were not present originally), although with a drop of only a few millibars in the general level of pressure.

5.6. With respect to the seasonal variation (Table 4b), the mean surface-pressure at the centre of the pool was much less in spring and summer than in autumn and winter, and the pressure anomalies changed from positive in the latter to negative in the former seasons. This was entirely due to the pools over the sea for in the case of pools over the land there was no corresponding variation in the mean pressure anomalies. The figures for the associated surface system (mainly lows) showed the same seasonal effects, but with lower mean-pressures and bigger negative anomalies, except in autumn when most of the highs occurred (Table 2).

There was practically no latitudinal variation of mean surface-pressure anomaly at the centres of the pools, either over the land or over the sea. This was also the case for the mean surface-pressure anomaly at the centres of the associated lows; there were not enough data on cold highs to decide one way or the other.

5.7. The lowest central pressure of an associated low was 976 mb. and the highest 1022 mb., but nearly 90% of the cases were within 10 mb. of the average (viz., 1001·5 mb.). The corresponding extremes for associated highs were 1025 and 1040 mb., respectively. A comparison with the 40-year statistics of the frequency of occurrence of all lows and highs in particular localities, revealed that the cold lows were usually shallow for the area in question whereas the cold highs were well up to their usual intensity.

The frequency distribution of surface-pressure anomalies at the centres of the pools and of the associated surface-pressure systems is shown in Table 5. The anomalies for all the highs and most of the ridges are positive, while most of those for the lows are negative. The remaining types have both positive and negative anomalies, although positive predominate. The most frequent anomaly (all the data considered) is a small positive one.

5.8. Surface-pressure changes at the centres of the pools were usually small: the overwhelming majority were less than 10 mb. (rise or fall) per day, the average irrespective of sign being 4·5 mb/day (5·5 mb/day over the sea and 3·5 mb/day over the land). The greatest pressure rise found was 16 mb. in 24 hours (with an associated surface low) and the greatest fall was 23 mb. ("cold drop" type): both were over the sea. There was a definite tendency at all seasons, no matter what the associated pressure system, for the pressure at the centre of the pool to return to normal; this was however less marked for pools over the sea.

The pressure changes at the centre of an associated low or high were usually very small, most of them being less than or equal to 5 mb/day. The greatest 24-hour rise was 11 mb. (over the land) and the greatest fall was 9 mb. (over the sea); both were in a low. The average changes irrespective of sign were 4·5 mb/day, whether over land or sea. There was no noticeable tendency for these central pressures to return to normal. However for depressions, whatever the initial anomaly, the chances were slightly in favour of a rise rather than a fall of central pressure, although day-to-day changes were very erratic both in sign and magnitude.

/Table 4(a)

Table 4a.

Type of associated surface-pressure system		Average pressure: mb.				Average thickness: tens of feet.	
		Centre of pool		Centre of pressure feature			
		P	△P	p	△p	h	△h
L	Land	1011.0	-4.5	1004.5	-9.5	1709	-69
	Sea	1004.5	-10.0	999.5	-15.0	1743	-49
	Both	1007.0	-8.0	1001.5	-13.0	1730	-55
T	Land	1014.0	0.0			1773	-46
	Sea	1013.0	+0.5			1758	-47
	Both	1013.5	+0.5			1765	-47
H	Land	1028.0	+13.0	1033.5	+20.0	1726	-40
	Sea	1029.5	+10.0	1036.5	+18.5	1707	-73
	Both	1028.5	+12.0	1034.5	+19.5	1719	-53
R	Land	1021.0	+2.5			1734	-52
	Sea	1021.0	-1.0			1830	-20
	Both	1021.0	+2.5			1740	-50
C	Land	1016.5	+1.0			1737	-56
	Sea	1019.5	+3.0			1769	-42
	Both	1017.5	+1.5			1749	-51
S	Land	1018.0	+3.0			1714	-60
	Sea	1016.0	+1.5			1743	-57
	Both	1017.5	+2.0			1727	-59
TOTAL	Land	1016.2	+0.6	1009.2	-5.6	1726	-59
	Sea	1010.6	-4.5	1001.9	-12.8	1748	-48
	Both	1013.3	-1.8	1004.8	-9.9	1737	-54

The symbol △ represents the departure from the 40-year normal pressure at the centre in question, e.g., $\triangle P = P - \bar{P}$, is the departure from normal surface pressure ($\bar{P}$) at the centre of the pool. The number of cases in each class are given in Table 2 (last main row and column) and are not repeated here.

/Table 4b

Table 4b

	WINTER				SPRING				SUMMER				AUTUMN			
	P	△P	p	△p	P	△P	p	△p	P	△P	p	△p	P	△P	p	△p
Land	1018.0	+1.0	1008.0	-8.5	1016.0	+1.5	1005	-9.0	1013.5	0.0	1004.5	-9.0	1018.5	+0.5	1016.0	+2.0
Sea	1017.5	+4.5	1015.5	-1.5	1007.0	-8.0	1000	-15.0	1009.0	-5.0	997.5	-16.0	1023.5	+4.0	1033.0	+14.0
All	1018.0	+2.0	1009.5	-7.0	1011.0	-4.0	1001.0	-13.5	1011.0	-3.0	999.0	-14.5	1019.5	+1.5	1020.0	+4.5
	h	△h			h	△h			h	△h			h	△h		
Land	1688	-63			1705	-63			1786	-44			1736	-62		
Sea	1726	-44			1734	-57			1787	-41			1755	-53		
All	1697	-59			1723	-59			1786	-43			1740	-60		

The symbol △ represents the departure from the 40-year normal pressure at the centre in question, e.g. $\triangle P = P - \bar{P}$, is the departure from normal surface pressure ($\bar{P}$) at the centre of the pool. The number of cases in each class are given in Table 2 (last main row and column) and are not repeated here.

/Table 5

Table 5

Frequency distribution of surface-pressure anomalies at the centre of cold pools. Figures in brackets are for the anomalies at the centres of the associated pressure systems

Type of associated surface-pressure system	-35 to -31 mb.	-30 to -26 mb.	-25 to -21 mb.	-20 to -16 mb.	-15 to -11 mb.	-10 to - 6 mb.	-5 to -1 mb.	0 to +4 mb.	+5 to +9 mb.	+10 to +14 mb.	+15 to +19 mb.	+20 to +24 mb.	+25 to +29 mb.	+30 to +34 mb.	Total
L	1 (1)	2 (4)	2 (7)	3 (15)	13(16)	22(18)	15 (8)	13 (2)							71 (71)
T						2	5	15	1	1					24
H									2 (1)	4 (1)	1 (2)	1 (2)	(1)	(1)	8 (8)
R							2	5	5	1					13
C						1	6	10	7						24
S					1	3	5	13	8	2	1				33
TOTAL	1	2	2	3	14	28	33	56	23	8	2	1			173

/6.

6. Weather associated with cold pools

6.1. The weather within the area of the cold pool (not that associated with the surface-pressure system) was classified at 0300 and 1500Z*, on surface charts in accordance with the following scheme.

The state of the sky was classified as b if the weather were predominantly fine ($< \frac{1}{4}$ cover of cloud) over the entire area, bc if it were partly cloudy (between $\frac{1}{4}$ and $\frac{3}{4}$ cover), c if it were mainly cloudy ($> \frac{3}{4}$ cover) and o for a complete overcast. These categories were sometimes entered if there were isolated stations within the area in question with greater or less cloud than the prescribed limits. A note was also made of the type of precipitation (rain, hail, snow or sleet), if any, and whether it was reported predominantly as showers, intermittent or continuous precipitation, and its intensity (light, moderate or heavy). The occurrence of thunderstorms was also noted.

Obviously this procedure is not completely objective, but it represents the sort of judgment that a forecaster has to make continually, and, without attaching too much importance to individual results, should provide some worthwhile general indications. In studying the diurnal variation, the fact that the observations were scattered over a wide sector which the sun takes 6-7 hours to traverse, was ignored, especially as for some purposes the data were segregated into land and sea (i.e., mainly over Europe and the Atlantic, respectively) which reduces the longitudinal spread.

6.2. Table 6 shows the number of cases in the various cloud categories at 0300 and 1500Z, and for the different types of pool. A distinction is also made between cases with no precipitation of any kind within the area of the pool and those with some precipitation. The figures in brackets correspond to the pools similarly distinguished in Table 2, viz., the cold lows which formed in relatively high latitudes. In one case no observations were available at 1500Z.

There were relatively few cases of a complete overcast or of fine conditions: a much higher proportion were partly cloudy but most were cloudy. On the whole there were slightly more cases with than without precipitation, but within the various cloud groupings the proportion with precipitation increased from nil with fine conditions to 100% with overcast, the figures for partly cloudy and cloudy being 45% and 57%, respectively. The L-type had the highest proportion with precipitation; the S- and C-types the lowest. The H- and R-types showed a surprisingly high proportion with precipitation. This was however invariably slight snow or sleet in the case of highs; but with ridges in association all forms and intensities of precipitation occurred, including one case of widespread moderate rain and another with thunderstorms. However, as previously stated (section 5.2), an R-classification was often one with weak pressure gradients, scarcely distinguishable from a C-type.

There was a slightly though not significantly higher proportion with precipitation at 0300Z than at 1500Z, especially in cold lows. In cases with no precipitation there was a definite decrease of b and bc from 0300 to 1500Z and a corresponding increase of c: with precipitation the tendency was for the o-type to decrease in the afternoon.

6.3. The diurnal variation indicated by the above table was studied in more detail, with data for land and sea segregated. The results are summarized below.

With respect to cloud, there was a definite diurnal variation over land in the sense of more cloud in the afternoon (there were about 55% occasions with o or c at 0300, and 70% at 1500Z). However, within this general trend from lower

/Table 6

* Since the coverage of observations is usually better at the main synoptic hours, the weather at 0300 was based on a general impression of the 0000, 0300 and 0600Z charts; and that at 1500 on the 1200, 1500 and 1800Z charts.

Table 6.

Pressure System	Time	o		c		bc		b	Total		Grand Total
		I	II	I	II	I	II		I	II	
L	0300	0	7 (1)	10 (4)	28 (8)	11 (2)	13	2	23 (6)	48 (9)	71 (15)
	1500	0	7	20 (5)	24 (8)	9 (1)	10 (1)	1	30 (6)	41 (9)	71 (15)
T	0300	0	3	5	6	3	2	5	13	11	24
	1500	0	3	6	6	3	4	2	11	13	24
H	0300	0	0	2	3	2	0	1	5	3	8
	1500	0	0	1	2	3	2	0	4	4	8
R	0300	0	2 (1)	3	3	2	3	0	5	8 (1)	13 (1)
	1500	0	0	4	6 (1)	1	1	1	6	7 (1)	13 (1)
S	0300	0	2	7	10 (1)	6	2	6	19	14 (1)	33 (1)
	1500	0	1	12	9 (1)	6	2	2	20	12 (1)	32 (1)
C	0300	0	3	6	4	6 (1)	2	3	15 (1)	9	24 (1)
	1500	0	1	7	7 (1)	2	3	4	13	11 (1)	24 (1)
Total	0300	0	17 (2)	33 (4)	54 (9)	30 (3)	22	17	80 (7)	93(11)	173(18)
	1500	0	12	50 (5)	54(11)	24 (1)	22 (1)	10	84 (6)	88(12)	172(18)
Grand Total		0	29 (2)	83 (9)	108(20)	54 (4)	44 (1)	27	164(13)	181(23)	345(36)

I. Without precipitation of any kind.

II. With precipitation.

/to

to higher cloudiness throughout the day there was a definite counter-drift (about 15% of the total changed from o or c at 0300Z to bc or b at 1500Z). There was no discernible diurnal variation in the cloud over the sea but this is more uncertain owing to the rather scanty observational coverage. The above remarks apply equally to the 6 classes of pools with different surface-pressure systems in association.

With respect to precipitation, there were slightly fewer cases with precipitation at 1500 than at 0300Z, both over land and sea, but the difference was scarcely significant (see Table 7, in which the sort of precipitation - showers or otherwise - is also classified). There was however a distinct tendency for rain to become more showery in the afternoon, although this may not be real owing to the difficulties of night-time observation. The figures in brackets are the cold lows in high latitudes as in Table 2. Roughly the same proportion of pools had precipitation at all seasons.

Table 7.

Precipitation in cold pools

Type of precipitation	Land		Sea		Totals	
	0300	1500	0300	1500	0300	1500
Rain or snow	35(1)	29(1)	23(4)	20(6)	58(5)	49(7)
Showers	13	15	22(6)	24(5)	35(6)	39(5)
Totals	48(1)	44(1)	45(10)	44(11)	93(11)	88(12)

6.4. A crude numerical measure of the intensity of precipitation was adopted, intermittent precipitation and showers being given less weight than continuous precipitation, and slight less weight than heavy. This was used as a basis for the following statements, which are perhaps more justified therefore than if they depended on a mere impression of the data.

The precipitation was on the whole slightly more intense over the sea than the land; bearing in mind the paucity of observations over the sea this may not however be a significant effect. There was very little change of intensity from pools with one kind of associated pressure system to another, with the exception of the H-type where the precipitation was invariably slight and intermittent. There was a slight diurnal variation of intensity, the higher values both over land and sea occurring at 1500Z. There was also a slight seasonal variation of intensity with a maximum in summer and a minimum in winter, but the fewness of the data in the latter season makes the significance rather dubious.

The association between intensity of precipitation and the surface-pressure at the centre of the pool was small (although in the sense of more intense precipitation the lower the pressure). However, in the case of associated lows a high proportion of pools without precipitation had relatively high surface-pressure: the lows were either shallow or well-removed from the pool.

6.5. Of the 62 spells, six were predominantly dry throughout with variable cloud (average duration 2.1 days), 13 were predominantly cloudy or overcast with fairly general precipitation most days (average duration 3.2 days; this figure is heavily weighted however by one eight-day spell), the remainder (average duration 3.0 days) being mainly fair to cloudy with intermittent and/or local precipitation, often infrequent. General precipitation over the entire area of the pool was recorded on about 20% of the total occasions, 30% had only local precipitation, and the remainder none at all.

/6.6.

6.6. Only six cases of thunderstorms were noted within the pools, one each over land and sea at 0300Z (L-type); two at 1500Z over land and one at 0300 over the sea (T-type); and one at 0300Z over the land (R-type). On the other hand there were very many more occasions of "sferics" reported within the area of the pool, mostly occurring in the warmer months over Europe; these showed a definite diurnal variation with the greater frequency in the afternoon. The positioning error in sferic-observations is however such that they might often have referred to thunderstorms in the peripheral regions of the pools. Although it is possible that some storms escaped observation, their reported infrequency within cold pools is probably real. The fewness of thunderstorms in the central regions of cold lows has been remarked on by E.A. Amman (1950).

6.7. There was a fairly clear-cut division in the values of the central thickness of the pool as between rain and snow. For showery precipitation, all cases of snow had a central thickness of 17,150 feet or below, and all cases of rain had a value above 16,900 feet. For precipitation other than showers (including intermittent as well as continuous precipitation) there was a fairly distinct separation at 17,200 feet; there were no instances of rain below this value and only three cases of snow (out of 47) with a higher value, none of them occurring with a thickness above 17,350 feet. There were only 6 cases of sleet (2 showers, 4 otherwise) and these had a central thickness between 16,950 and 17,200 feet. The intensity of the precipitation did not affect these criteria perceptibly.

Figs. 6a, b and c show the frequency distribution of central thickness, rain, snow and sleet being distinguished. Both the 0300- and 1500- observations are taken into account in these figures - and in the values given in the above paragraph: the implication is that if the thickness changed across any of the above limits between these times, the nature of the precipitation would also change.

7. 1000-500 mb. thickness values at the centres of cold pools

7.1. Tables 4a and 4b include the average central thickness (h) and thickness anomaly (Δh; departure from the 5-year mean), for each associated surface-pressure system and for each season, respectively, both on a land-sea basis.

The average central thickness shows a marked seasonal variation, and except for summer it is significantly less over the land than the sea; all these differences are in the sense one would expect. The land-sea difference also shows up in most of the different types of cold pool (Table 4a).

The average thickness anomaly for all cases was -540 feet, -480 feet for pools over the sea and -590 feet for those over the land. The numerical value of the anomaly was on the whole smallest in the summer, but there was also another minimum in winter in the case of pools over the sea (the number of cases is however small). The difference in anomaly between land and sea was maintained throughout the seasons, but not for all the individual types (Table 4a).

7.2. The average anomaly was (numerically) a few feet more at the commencement of a spell than at the end, the general tendency being for a pool to warm a little with time (see section 7.5). There was no relationship between the initial anomaly and the subsequent duration of a spell.

There was a marked and smooth latitudinal variation of mean-thickness anomaly. The data were segregated in 5° latitude bands from 30° to 65°N, and the greatest departure from normal (-660 feet) occurred in middle latitudes (46-50°N) falling fairly uniformly to minima on either side (anomaly about -375 feet in both cases). This sort of distribution was in evidence both over land and sea.

/7.3.

7.3. The scatter of the individual values of the anomalies about the mean was not very large: standard deviations were not computed, but from an inspection of the individual data in each category it seemed that values of about 30-40% of the mean would be the greatest one would expect. The greatest and least anomalies for all the data were -1,150 feet and -150 feet, respectively, but such extreme values were very infrequent. (As a point of interest, every pool had a central thickness below the 5-year mean for the time and place, and for as long as it maintained the requisite intensity).

7.4. In general the larger initial anomalies were followed by rising central thickness and the smaller anomalies by falling thickness. Fig. 7 is a dot-diagram of the central thickness anomaly (abscissa; mean of values at beginning and end of 24-hour periods) against the 24-hour change of thickness (ordinate). There is a rough relationship; the line of best fit, judged by eye, cuts the X-axis at about -450 feet anomaly. Since we are following the changes which occur at the centre of the pool, thermal advection may be neglected, and the implication is that warming from below and dynamical and radiational processes (presumably a net cooling) tend to balance out about this value of the thickness anomaly.

In Fig. 7 the cases with fairly widespread precipitation are distinguished from those with only local precipitation and those with none at all. Except for some reduction in the scatter, there seems to be little difference between the three classes. Other segregations were tried such as into seasons, into land and sea, or into the various classes with different surface-pressure systems associated. In every class containing sufficient cases to judge from, there was little to distinguish them from the main body of data except for a slight reduction in the scatter and a small shift of the best-fitting line.

There was no relationship between 24-hour surface-pressure and thickness changes at the centres of the pools.

7.5. There was a definite diurnal variation of central thickness with the higher values in the afternoon both over land and sea, the average amplitude being about +75 feet and +60 feet per day, respectively. These values were based on the average difference of 1000-500 mb. thickness between the 1500 and 0300Z values, with the mean "trend" (amounting to about +40 ft/day over the land, and +35 ft/day over the sea) eliminated.

The diurnal variation in the winter months was only 50 feet over land, but 100 feet over the sea; however there were only 9 cases of the latter and the increase as compared with the figure for all the data (viz., 60 feet) is scarcely significant. A selection of pools with broken cloud (b or bc) at one or both of the 0300 and 1500Z observations and with no precipitation at either, showed very similar values for the diurnal variation as for all the data. From an inspection of the data there was no reason to suppose that any other selection would give very different results.

It may be assumed that the diurnal variation is due to the interaction between the two factors: (a) the net long-wave radiational heat loss from the cloud and the air itself, going on night and day without much change; and (b) eddy fluxes of heat into the column (resulting from temperature changes at the undersurface) and the absorption of short-wave radiation by water-vapour and clouds, both due to insolation and therefore having a diurnal variation with a maximum in the afternoon (this would be most marked in the warmer months when most of the cases occurred). With this oversimplified picture, then, on the assumption that (b) operates for half the day (or more) and that the values found represent the total range, the radiational loss from the 1000-500 mb. layer (factor "a") is roughly about 2°C per day (or less) overland, and somewhat lower over the sea. This is in keeping with other estimates currently accepted.

/8.

8. Concluding remarks and forecasting application

8.1. The above data refer only to a particular area of the northern hemisphere and to cold pools of a certain minimum intensity. Their forecasting application, although perhaps not actually confined to similar cases, is certainly most appropriate to them. The forecaster is always interested in the limits within which he can work and these data may be of use in providing some of this background knowledge. But his chief concern is with individual cases, which were not studied here in the detail which is necessary (and usually available) in day-to-day forecasting. However, arising from our statistics there are a few rules and general indications, summarised below, which may serve as a more specific guide in synoptic practice.

8.2. (a) Cold pools comprising two or more closed 1000-500 mb. thickness lines, once having formed (appeared on an 0300 Z chart) are almost invariably present 24 hours later. The average persistence is 3 days, i.e. on an average they appear on three sucessive 0300 Z charts at this intensity.

(b) Although present at all seasons, they are mainly late spring and early summer phenomena.

(c) Intense cold pools are usually slow-moving in all latitudes; in particular, pools north of 65°N do not move south of this latitude.

(d) Except for a small number of cold lows in latitude 55-65°N (all in the northern Atlantic and in the summer months), most intense cold pools are formed in lower latitudes as a result of a cut-off process (illustrated in Fig. 5), and associated with a blocking of the westerlies. There are, however, many variants of this basic model especially in respect of surface developments.

(e) Intense cold pools are very unusual on the eastern Atlantic in the latitudes of the British Isles, near Iceland (all seasons) and in the south-west Atlantic in summer (see Figs. 1 and 2).

(f) The central thickness of pools of intensity two is invariably below - usually 400-500 feet below - the "normal" for the time and place, with some tendency to "relax" to an anomaly of about -450 feet if greater or less than this value.

(g) Most pools lose intensity slowly, the most important factor in their dissipation being warming in situ (i.e. a raising of the central thickness). About half the pools of intensity two subsequently persist for at least a day as a pool enclosed by one closed thickness line before finally disappearing from the charts.

Colder air from the north seldom breaks through again into a cut-off cold pool.

(h) Cold pools may be associated with any type of surface-pressure distribution, but more commonly with a (fairly shallow) surface depression, especially when situated over the sea (Table 2). Cold highs are rare, and are practically non-existent in spring and summer. There is a definite tendency for the straight isobars associated with a "cold drop" to become cyclonically curved.

(i) The 24-hour change of surface-pressure at the centre of cold pools and at the centre of an associated surface-system is usually well below 10 mb.

/(j)

(j) Precipitation occurs with all types of cold pool, no matter what the associated surface-pressure distribution, but is more frequent in cold lows than in any other type.

Thunderstorms are surprisingly infrequent in the central regions of cold pools.

(k) Most intense cold pools have more than $\frac{3}{4}$ cover of cloud, although a complete overcast is rare - in fact just as rare as mainly fine (Table 6).

(l) There is a slight diurnal variation of central thickness (higher value at 1500Z) in pools situated both over the sea and the land. This averages about 60 feet over the sea and 75 feet over land.

(m) Precipitation is in the form of snow with a central thickness below 16,800 feet, and of rain with a value of 17,400 feet or above. For showers this latter value is 17,200 feet, and for more widespread (although sometimes intermittent) precipitation the former limit is 17,000 feet.

Useful working values, which in each class consitute a fairly clear-cut demarcation (Fig. 6), are 17,000 feet for showers and 17,200 feet for more general precipitation.

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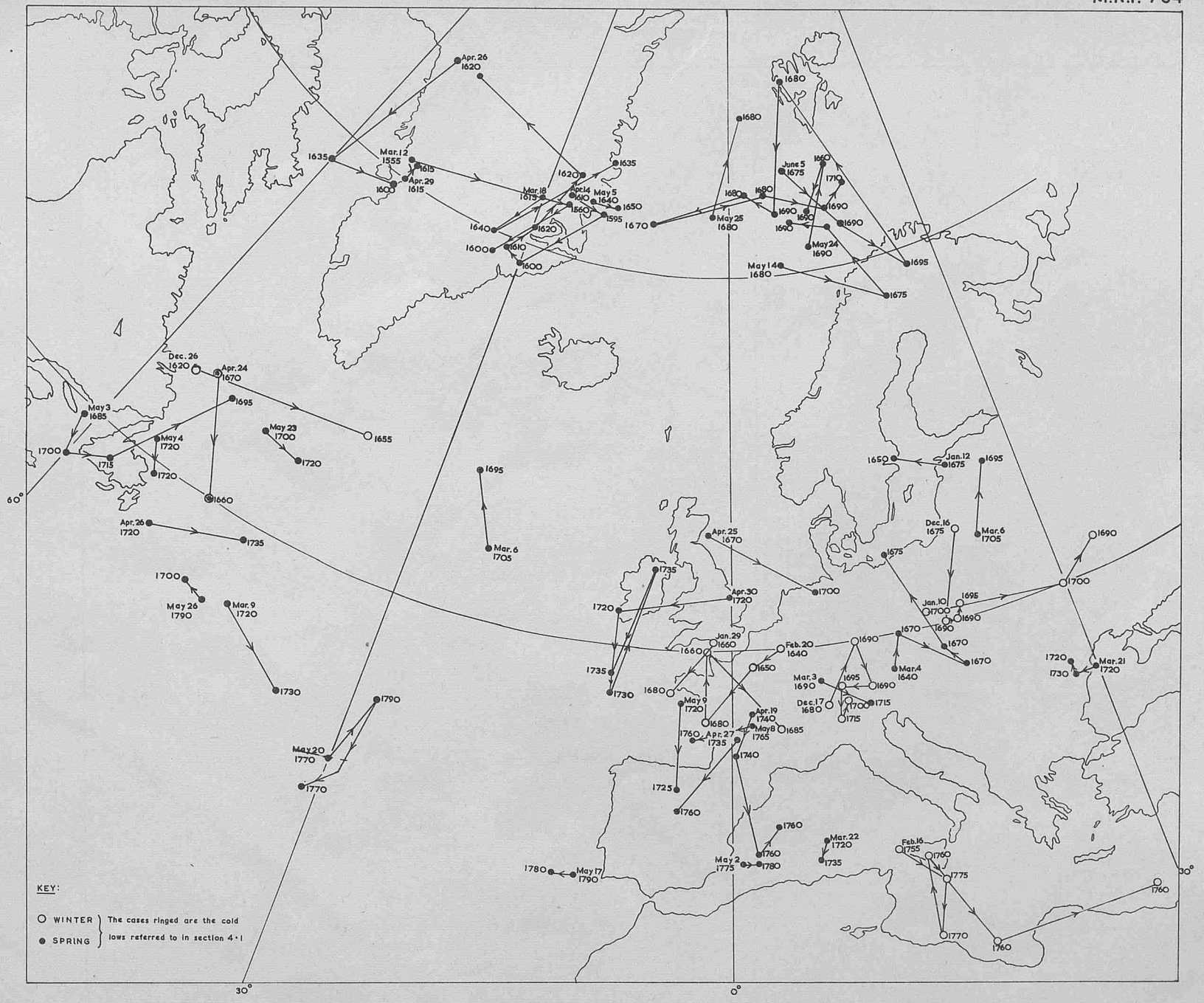


Fig.1 Position, movement and central thickness of cold pools: winter and spring.

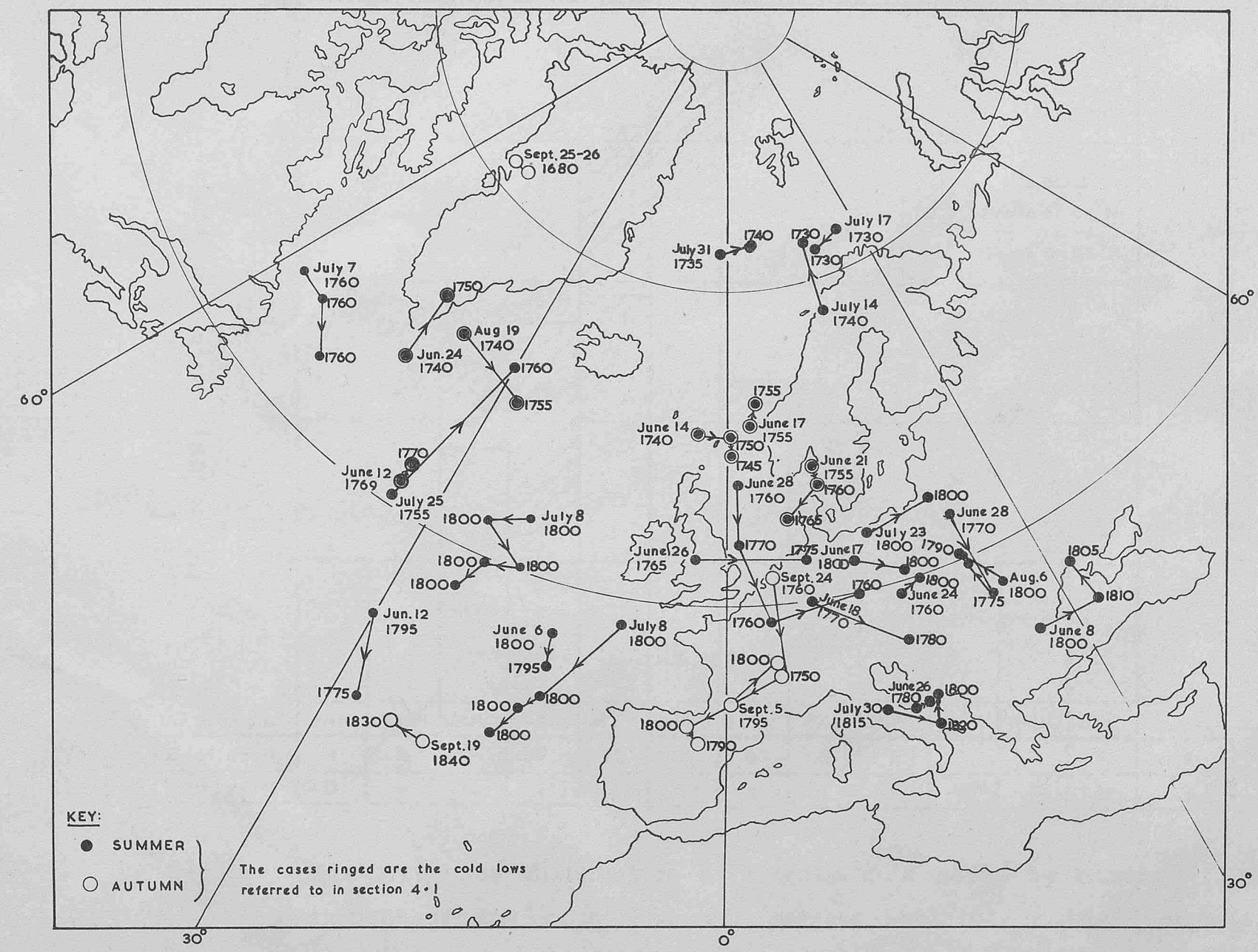
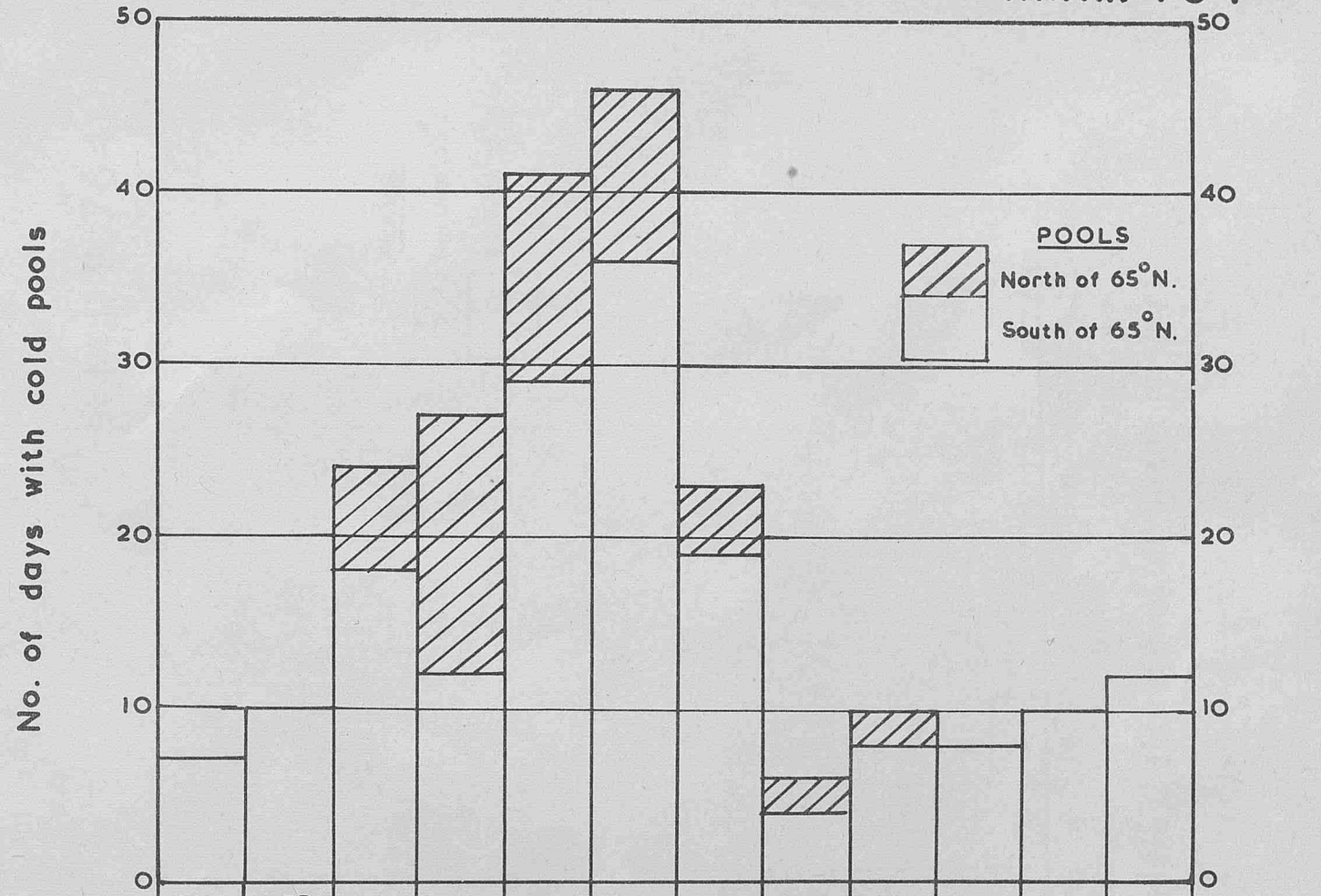


Fig. 2 Position, movement and central thickness of cold pools: summer and autumn.



No. of spells	JAN.	FEB.	MARCH	APRIL	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	Total
S. of 65°N	3	2	7	5	12	14	6	2	3	3	2	3	62
N. of 65°N	0	0	2	2	4	1	2	1	1	0	0	0	13

FIG. 3 The frequency distribution of intense cold pools by months throughout the year. Five year period Sept. 1946 - Aug. 1951., and within the area between longitude 60°W to 30°E and south of latitude 80°N.

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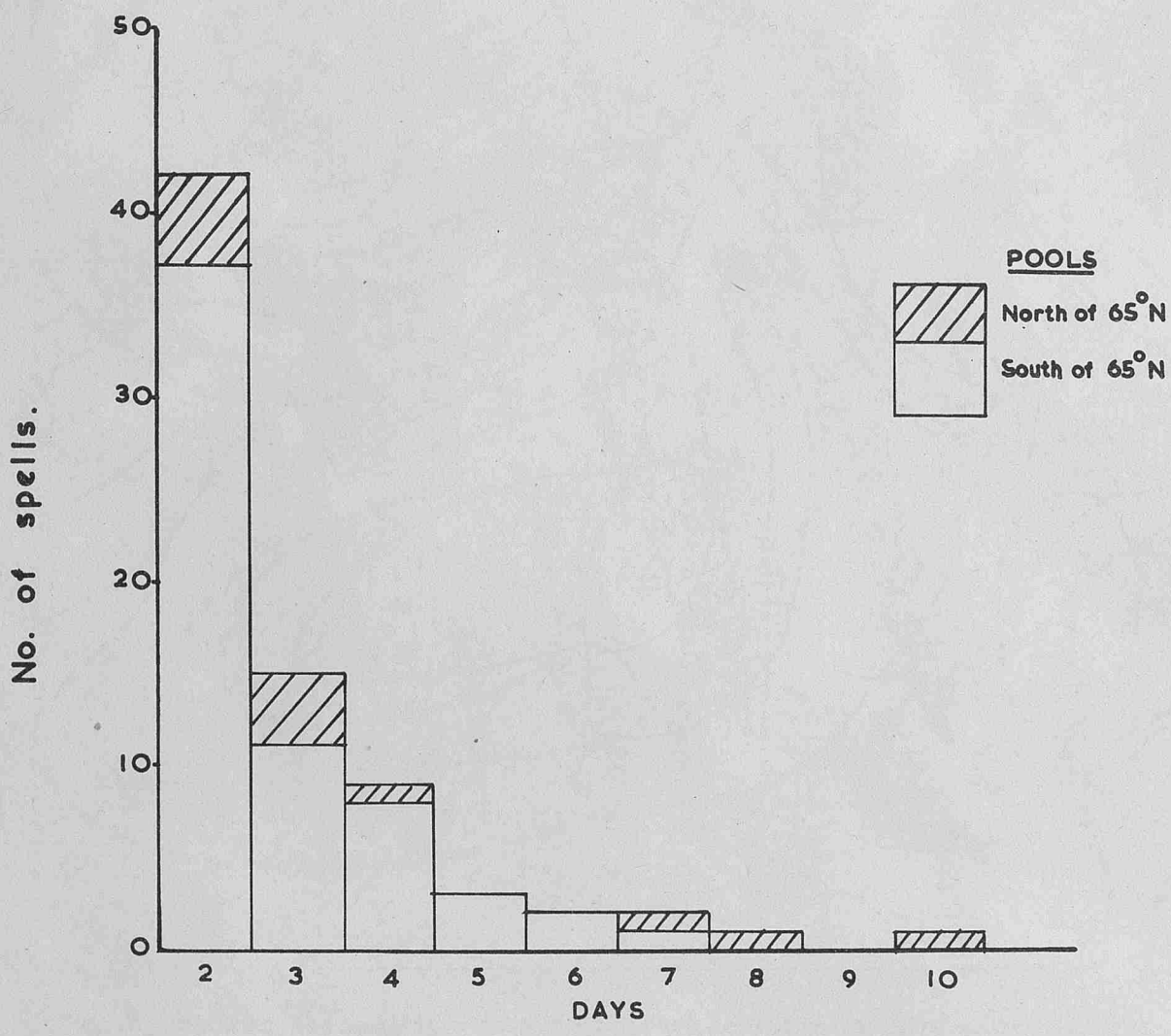
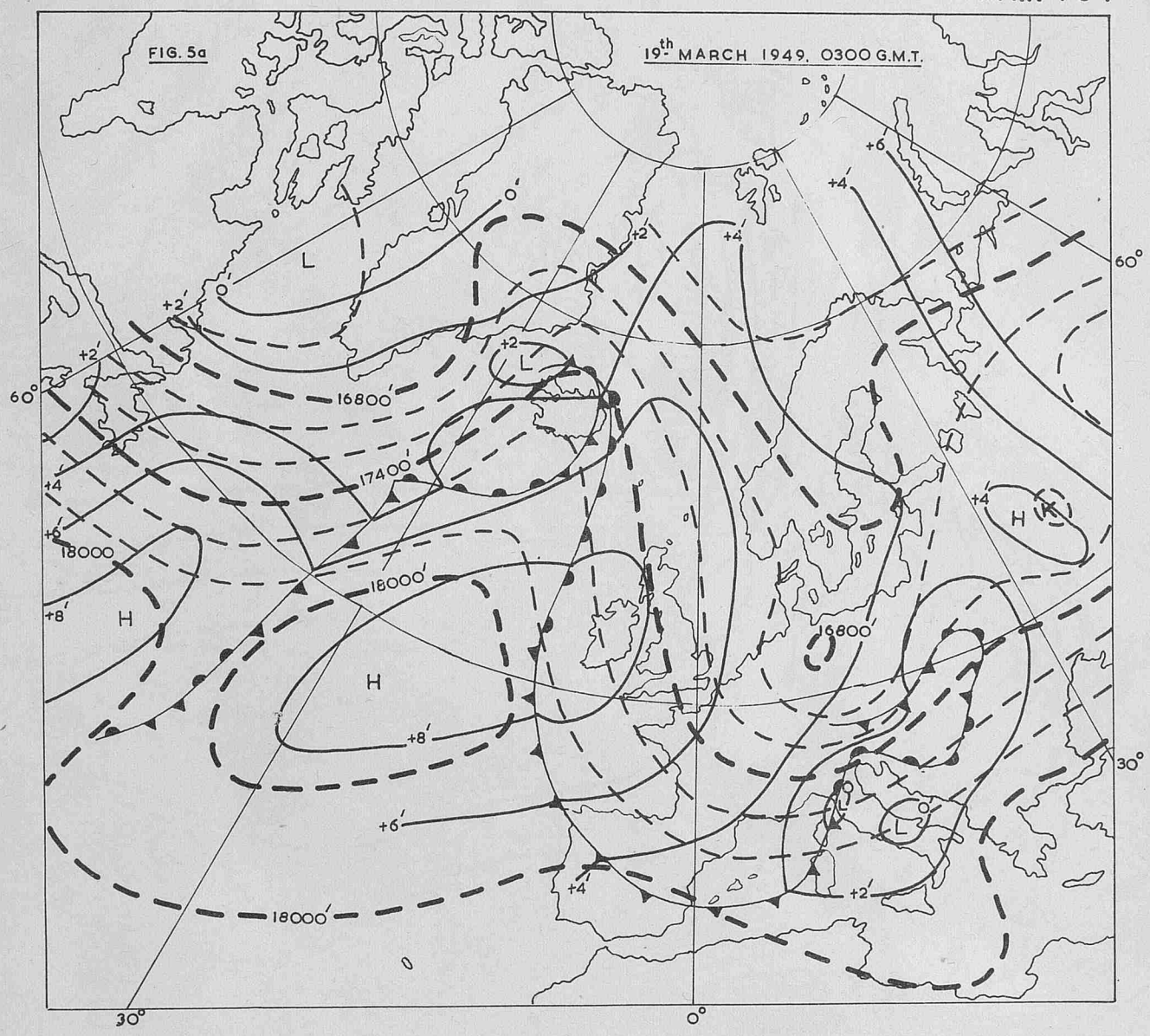
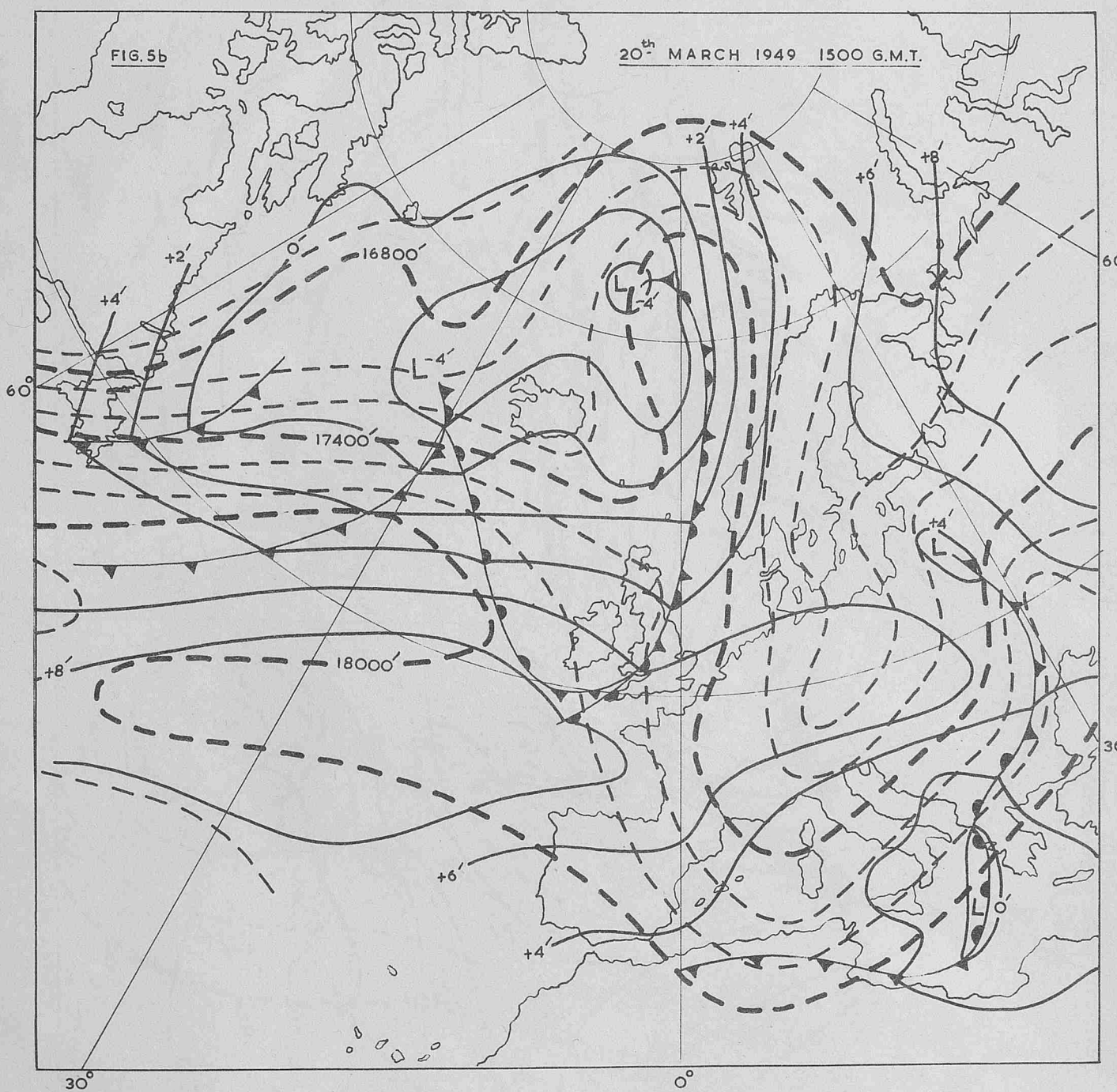


FIG. 4 The frequency distribution of spells in days.



Figs. 5 Showing the cutting off of a cold pool near the southern extremity of a cold trough.



1000–500 mb. thickness lines marked in feet at 600 ft. intervals.

Intermediate thickness lines at 200 ft. intervals.

1000 mb. contour heights marked in hundreds of feet, at 200 ft intervals.

The surface fronts are marked in the usual symbols.

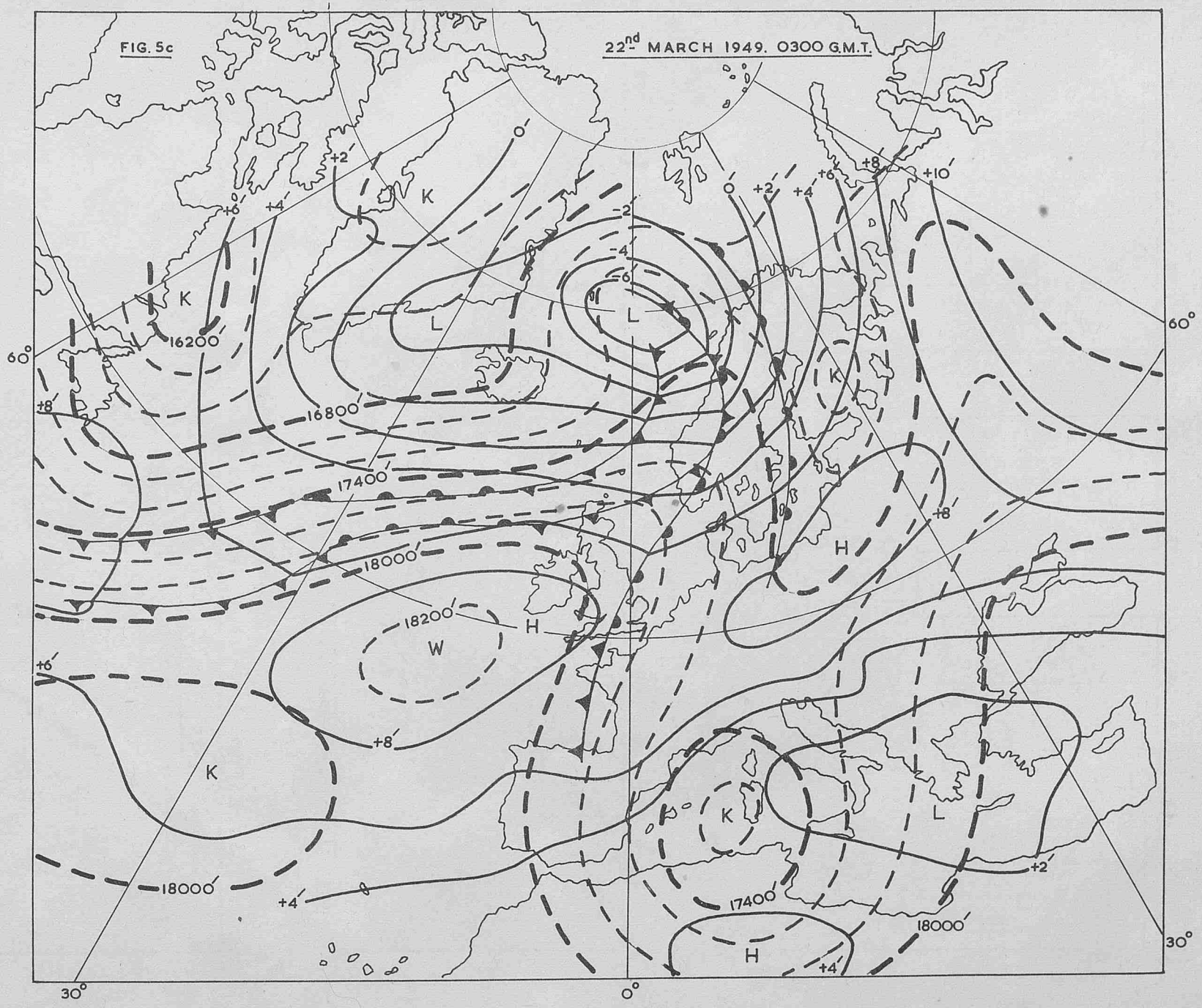


FIG. 5c
22nd MARCH 1949. 0300 G.M.T.
60°
30°
0°
0'
+2'
+4'
+6'
+8'
+10'
-2'
-4'
-6'
16200'
16800'
17400'
18000'
18200'
K
L
H
W

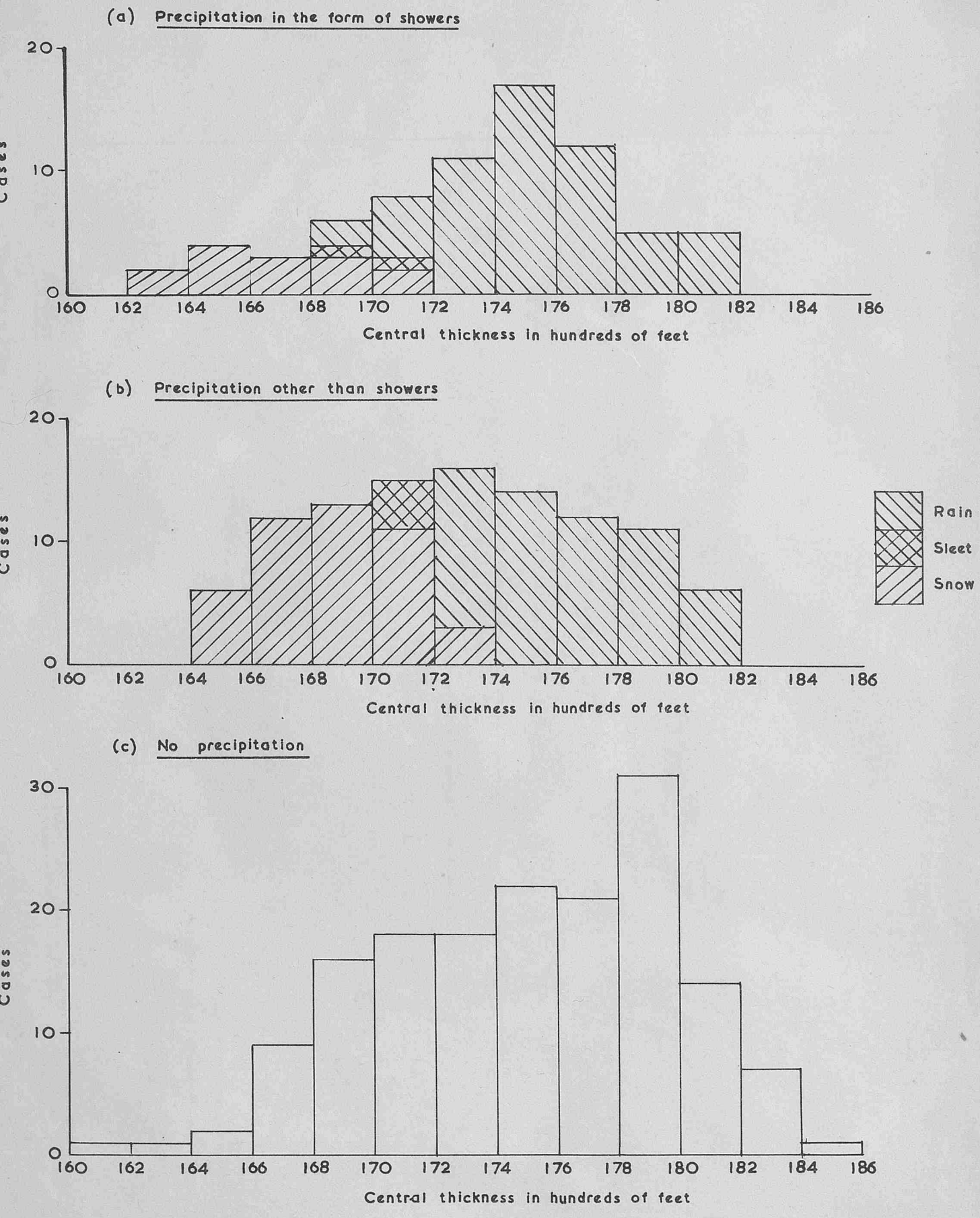


FIG. 6 Frequency distribution of 1000-500 mb. thickness at the centres of cold pools.

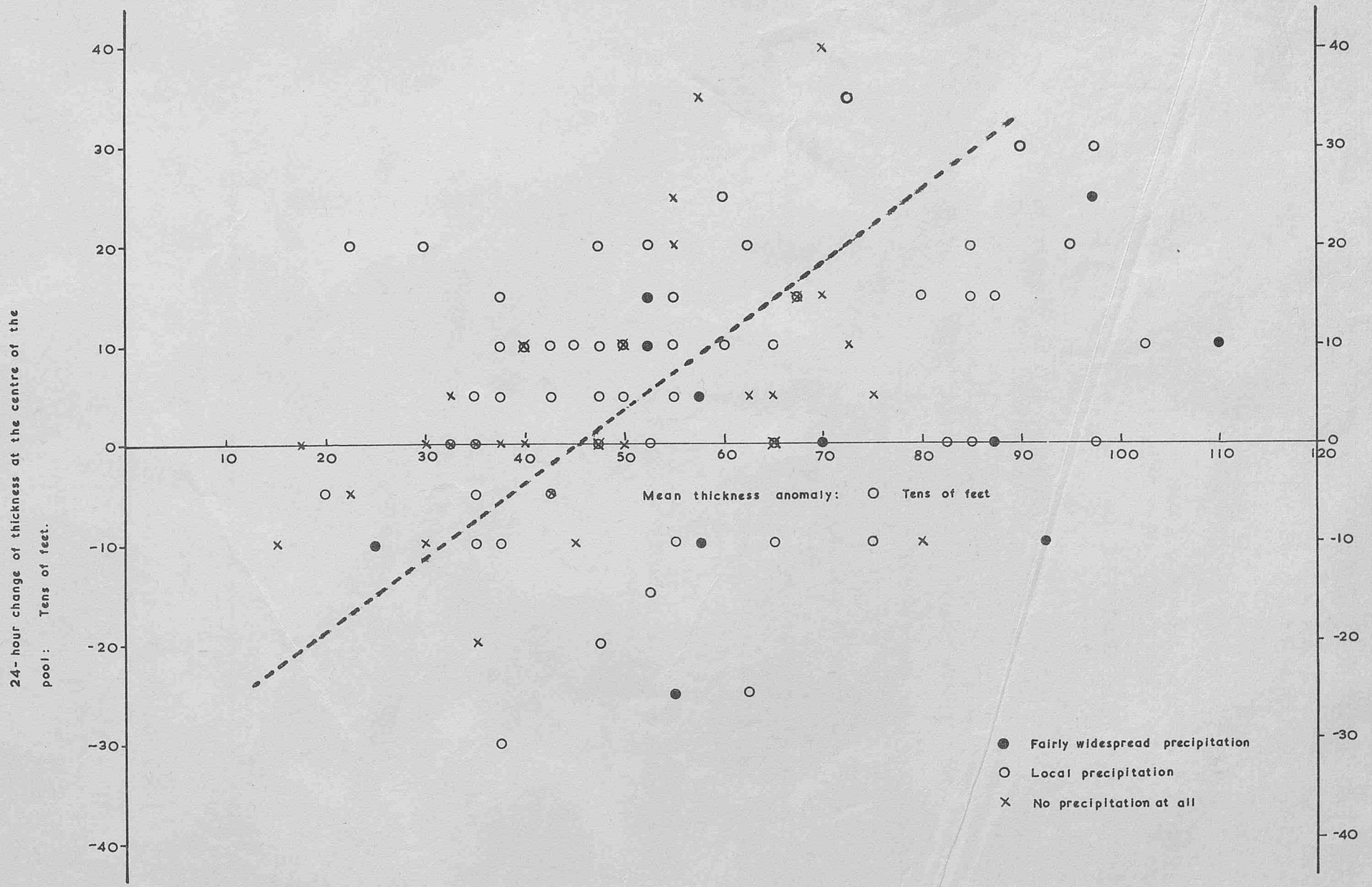


Fig. 7 24-hour changes of central thickness in relation to the mean thickness anomaly

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M.R.P. 765
S.C.II/126
8th January, 1953.

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METEOROLOGICAL RESEARCH COMMITTEE

COMPUTATION OF THE FIELD OF ATMOSPHERIC DEVELOPMENT BY AN ELECTRONIC COMPUTER

by
F.H. Bushby and Mavis K. Hinds (Meteorological Office)

Introduction.

Sutcliffe (1), after various approximations, derived an expression for cyclonic or anticyclonic development in terms of rates of change of vorticity and the vertical shear of the horizontal wind field. This formula has been used qualitatively in the Forecast Division at Dunstable for 3-4 years and Sutcliffe and Forsdyke (2) have shown how the principle can be used in medium range forecasting. Although the Sutcliffe development term can be calculated by hand computation it is a laborious process, even when the Sawyer and Matthewman (3) scale is used. Consequently there has been little attempt to investigate the merits of computing the field of development and using it as an aid to forecasting.

The advent of modern electronic computing machinery has meant that the time factor involved in calculating the field of development is no longer a deterrent to its use by the forecaster. As there seems little doubt that a qualitative assessment of the maximum areas of development by inspection of the chart is of use to the forecaster, there would seem to be an "a priori" case for having the field of development computed on a routine basis. The present report gives a brief description of a method by which the field of development can be calculated by an electronic computer. The results of 12 such computations are discussed in relation to the relevant synoptic situations.

Essentially the method consists of evaluating a finite difference approximation to a Jacobian at a number of grid points. A similar type of computation is necessary as an intermediate step in the solution of the differential equations derived by Bushby and Sawyer (4), and the experience gained in preparing the present programme will be invaluable in preparing more complicated programmes.

The Sutcliffe development formula.

Sutcliffe (1) has shown that, subject to various approximations, the relative divergence between two pressure levels p_0 and p_1 is given by

$$l\left(\mathrm{div}_p \underline{V}_1 - \mathrm{div}_p \underline{V}_0\right) = -\underline{V}' \cdot \mathrm{grad}_p\left(l + \xi_0 + \xi_1\right) \qquad \ldots\ldots(1)$$

where l is the Coriolis parameter; $\underline{V}_0$, $\underline{V}_1$, ξ_0 and ξ_1 are horizontal wind velocities and the vertical components of vorticity at the pressure levels p_0 and p_1 ; $\underline{V}'$ is the vertical wind shear $\underline{V}_1 - \underline{V}_0$, and the suffix p represents differentiation in an isobaric surface with respect to orthogonal curvilinear distance co-ordinates on the sphere.

Sutcliffe (5) has shown how the left hand side of Eq. (1) can be interpreted in terms of vertical motion and that geostrophic values may be used in evaluating the right hand side. The equation is particularly valuable because

the horizontal divergence which appears on the left hand side cannot be estimated from geostrophic winds, and even when the most dense network of actual wind observations is available the values of divergence which can be obtained are inadequate to give a satisfactory estimate of the difference of divergence between two levels. It is, however, practical to compute the right hand side of Eq. (1) and this gives an approximate method for computing the field of vertical motion.

In this connection it should be noted that Sumner (6) and Bushby (7) have shown that although the omission of the vertical stability term from Eq. (1) may seriously affect the magnitude of the computed vertical velocity it is unlikely to affect the positions of the areas of maximum and minimum vertical velocity.

By using the geostrophic approximation, we obtain

$$\underline{V}' \cdot \mathrm{grad}_p(\ell + \xi_0 + \xi_1) = g\ell^{-1} J_p\{z', \ell + g\ell^{-1}\nabla^2(2z_1 - z')\} \quad \ldots\ldots(2)$$

where z_1 is the height of the isobaric surface p_1 ; z' is the thickness of the layer from p_0 to p_1 ; g is the acceleration due to gravity, and J_p and ∇_p^2 are the Jacobian and two-dimensional Laplacian operators, referred to orthogonal curvilinear distance co-ordinates.

Conformally mapping the northern hemisphere on to a plane by a stereographic projection from the south pole on to a plane perpendicular to the axis of the earth passing through the north pole, the right hand side of Eq. (2) transforms to

$$m^2 g\ell^{-1} J\{z', \ell + m^2 g\ell^{-1}\nabla^2(2z_1 - z')\} \equiv D \quad \ldots\ldots(3)$$

where m is the magnification factor $\sec^2\left(\frac{\pi}{4} - \frac{\theta}{2}\right)$, θ being the geographical latitude, and J and ∇^2 are the Jacobian and two-dimensional Laplacian operators referred to rectangular Cartesian co-ordinates.

The finite difference approximations used in evaluating D (defined by Eq. (3)) at a network of points were

$$[J(\alpha,\beta)]_0 = \frac{1}{4h^2}\left[(\alpha_1 - \alpha_3)(\beta_2 - \beta_4) - (\beta_1 - \beta_3)(\alpha_2 - \alpha_4)\right] \quad \ldots\ldots(4)$$

$$[\nabla^2\gamma]_0 = \frac{1}{h^2}\left[\sum_1^4 \gamma_1 - 4\gamma_0\right] \quad \ldots\ldots(5)$$

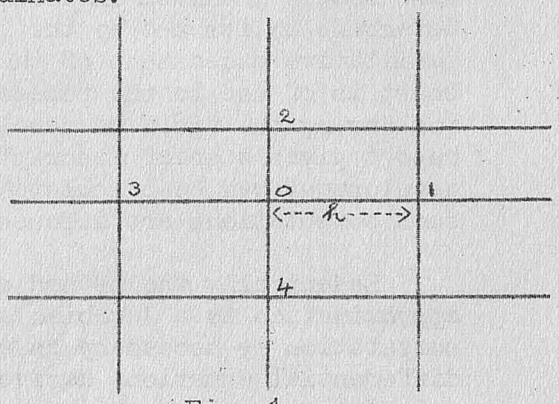

Fig. 1.

where the suffixes refer to the grid points of Fig. 1. The grid length used at 90°N was R/20, where R is the radius of the earth.

The Computation.

The programme MET 2 for computing D on LEO at a network of 12 x 8 points occupies 281 storage locations. By changing the pre-set parameters of the sub-routines the programme can be easily adapted to compute D at any network of r x s points, whatever the grid length. (3rs + 12s + 10 r + 34) storage locations are needed to store the data.

The time taken by LEO to compute D on a 12 x 8 network of points was $5\frac{1}{2}$ minutes. 3 minutes were spent in reading in the programme and data, 1 minute on the calculations and $1\frac{1}{2}$ minutes on printing the results. It is necessary to allow a certain time for reading data off charts, punching data tapes and plotting the answer, but it is estimated that a development chart could be produced in 1 hour from the time the initial charts were ready.

This compares with 4-5 hours when the most efficient method of hand computation (Sawyer and Matthewman (3) scale) is used, and no corrections for variations of the Coriolis parameter and scale of the chart with latitude are made in the hand method.

To ensure that the programme had been prepared correctly, one computation was done by hand and all the corrections were incorporated. Including checking, 24 hours were spent on this. This computation was used as a check and was run at the beginning, in the middle and at the end of the other 11 sets of data. An alternative check against random errors of LEO is to repeat each computation twice.

The Results.

Each development chart has been studied in relation to the synoptic situation. A sequence of surface charts with the 1000-500 mb. thickness pattern superimposed has been reproduced for each occasion that a development chart was computed. It should be remembered that cyclonic (anticyclonic) development is directly associated with upward (downward) motion, and only indirectly with surface pressure changes.

1500 G.M.T. 14th March, 1949.

The first computed development chart (Fig. 3b) was for 1500 G.M.T. on 14th March, 1949 and a sequence of three synoptic charts about that time is shown in Figs. 2a, 3a and 4a. Fig. 3b in conjunction with Fig. 3a indicates mobility of the main troughs and ridges in the latitude of the British Isles, as maximum anticyclonic (positive) development areas occur to the east of the main ridges and cyclonic (negative) development areas occur to the east of the main trough. Corresponding movements did in fact occur. A negative area between the Faeroes and Lerwick and a marked positive area slightly further west are in agreement with the formation of a cut-off cold pool to the north of Lerwick and the strengthening of the surface pressure ridge at 10°W. Another feature of note is the negative area about 30°W., 55-60°N. Quite a well-marked surface trough had appeared in this region by 0300 G.M.T. 15th March with a break-away depression at 63°N. 30°W. just off the edge of our chart.

Since the Sutcliffe development term is directly proportional to the mean vertical velocity in the model used by Bushby (8) when the vertical stability term is neglected, the isopleths on Fig. 3b should be the same pattern as those of the undamped vertical velocity for the same time shown in reference (8). The fact that the patterns are in close agreement shows that the differences in technique in using the Sawyer-Matthewman (3) scale or in evaluating the appropriate Jacobian do not seriously affect the development patterns. This point is of special interest as the grid length used in the present computations was about 160 miles in the latitude of the British Isles, whereas the grid length of the Sawyer-Matthewman scale was only about 100 miles.

1500 G.M.T. 19th October to 0300 G.M.T. 21st October, 1950.

A sequence of 4 development charts (Figs. 6b, 7b, 8b and 9b) at 12 hourly intervals was computed for the period from 1500 G.M.T. 19th October to 0300 G.M.T. 21st October, Appropriate synoptic charts are shown in Figs. 5a, 6a, 7a, 8a, 9a and 10a. This situation was chosen as it shows the evolution of a slow moving anticyclone in the vicinity of the British Isles from a mobile Atlantic ridge. The main increase in intensity and amplitude of the anticyclone took place between mid-day on the 19th and mid-day on the 20th. The amalgamation of the anticyclone over S.E. Greenland with the Atlantic ridge took place outside the area of the development charts, but nevertheless there are several features of interest in the computed charts.

The development chart (Fig. 6b) for 1500 G.M.T. on the 19th October shows a maximum area of anticyclonic development N.E. of the Atlantic cell, with another pronounced maximum NNW of Scotland behind the depression in the Norwegian sea. These two cells are only separated by a narrow tongue of weak cyclonic

/development

development, and it may be that with such a distribution of development one should look for an increase both in extent and intensity of the anticyclonic cell. In this connection it should be noted that when the vertical stability is included in the equation for vertical velocity a small minimum area of the Sutcliffe development term adjacent to a more intense maximum area frequently gives one large area of descending motion (Bushby (7)). Thus the development distribution of Fig. 6b might well imply an extensive area of subsidence both over and between the two aforementioned anticyclonic development areas.

The cyclonic development area orientated NW - SE on the western edge of the chart agrees quite well with the position of a surface trough 12 hours later, and in fact the analysis was changed at this stage and the warm front placed in this trough.

Over Europe the anticyclonic development area over Switzerland with a tongue stretching out over southern England and Wales and a cyclonic development area over Germany is certainly compatible with the slight re-orientation of the European anticyclone and the building across the British Isles. The cyclonic development area in the Bay of Biscay is not associated with any obvious surface development, but there is cooling in this area whereas simple advection would give warming.

At 0300 G.M.T. on the 20th, the maximum anticyclonic development area (Fig.7b) was situated over southern Norway, and extended into the north of the North Sea. The development pattern was then weakly anticyclonic to the west of the axis of the Atlantic ridge, indicating that cyclonic activity was unlikely to attack this flank of the anticyclone. The main cyclonic development area was farther west to the north of the warm front and this front did in fact move north. The weak cyclonic development area from Scotland to southern Iceland does not appear to be connected with any weakening of the ridge in that region.

The anticyclonic development area over south-eastern Germany was associated with the maintenance of high pressure in that region, while the cyclonic development area further south was associated with a fall of pressure. Another noteworthy feature was the cyclonic development area in the central Baltic, as 12 hours later the main cyclonic development was transferred from the Norwegian Sea to the Baltic.

At 1500 G.M.T. on the 20th October the main anticyclonic development was still situated over southern Norway, extending into the North Sea and to the north of Scotland. Over Scotland itself is a curious bulge of weak cyclonic development. This appears to be a feature which shows up the limitations of a method which relies on deducing the surface pattern from the difference between the 500 mb. height and the 1000-500 mb. thickness. However, if the vertical stability term were included, this feature would almost certainly be completely damped out by the large anticyclonic development area over southern Norway. At this stage there are still no large cyclonic development areas threatening the flanks of the Scottish anticyclone, although there is a general area of weak cyclonic development from Italy to southern Ireland. The anticyclonic development over Germany has now become less marked.

On the western edge of the chart the frontal trough is still indicated to be moving away northwards, as in fact happened. A large negative value north of the Azores would indicate a probable wave development in that region, whereas in fact none occurred. This is a difficult area of the chart to analyse and the computed value may be based on erroneous data.

At 0300 G.M.T. on the 21st October the computed anticyclonic development (Fig. 9b) over southern Scandinavia was only about half of its value 12 hours earlier, but pressure was still rising in this area. A small area of weak cyclonic development west of Denmark appears to be insignificant. There is still only a weak cyclonic development pattern to the west of the North Sea anticyclone, the main cyclonic development area being to the north of the main Atlantic depression. Apart from some curious features on the edge of the chart north of the Azores, there are no other significant development areas on this chart.

2100 G.M.T. 14th February to 0300 G.M.T. 15th February, 1950.

A well-marked warm front crossed the British Isles during this period and development charts (Figs. 12b, 13b) were computed for 2100 G.M.T. 14th and 0300 G.M.T. 15th February, 1950. The synoptic situation for this period is shown in Figs. 11a, 12a, 13a and 14a. The rainfall associated with this warm front has been discussed by Sawyer (9) and the field of vertical velocity has been evaluated by Bushby (7). A comparison between the field of undamped vertical velocity and the Sutcliffe development pattern again emphasises that the patterns obtained are independent of the techniques used.

The main cyclonic development area at 2100 G.M.T. on the 14th February was due east of the point of occlusion, and in fact the heaviest rainfall over the British Isles associated with this frontal system occurred over Scotland, although this was partly due to orographic effects. The secondary maximum of cyclonic development over N.E. France was also associated with a maximum of rainfall, although this rainfall had no apparent connection with the surface front. There was no appreciable surface pressure fall in this area.

Over England the weak anticyclonic development pattern is damped out when account is taken of the vertical stability but the computed vertical velocity over the Midlands and E. England, although upwards, is much less than over the rest of the British Isles. This agrees quite well with the rainfall distribution.

In the central Atlantic the main anticyclonic development pattern was behind the cold front, consistent with the rise of pressure immediately behind this front, The anticyclonic development pattern over southern France is associated with the maintenance and slight intensification of the high pressure ridge in that region.

At 0300 G.M.T. on the 15th February the cyclonic development area between 10° and 20°W. at 60°N. extends behind the occlusi n and pressure did fall in this area. Further south on the Atlantic there is a marked anticyclonic development pattern behind the cold front, apart from a small feature about 51°N. This would probably be damped out when the effect of vertical stability is considered, but it might lead one to expect a wave development in this region whereas none occurred.

In the vicinity of the British Isles there is a large area of cyclonic development ahead of the warm front, apart from some anticyclonic development in the centre of the North Sea. It does not seem that all this would be damped out by the adjacent cyclonic development and one would therefore deduce that there would be a certain region ahead of the warm front comparatively free from rain. It is impossible to verify or disprove this, as there is a scarcity of observations over the area concerned.

The maximum anticyclonic development over south-eastern France is associated with the strengthening of the high pressure ridge in that region.

0300 G.M.T. to 1500 G.M.T. 4th February, 1951.

Development charts (Figs. 16b, 17b) were computed for 0300 and 1500 G.M.T. on the 4th February, 1951 which was an occasion when an extremely deep depression moved westwards from the Atlantic across Ireland and into the Irish Sea (Figs. 15a, 16a, 17a and 18a). At 0300 G.M.T. on the 14th there was one strong cyclonic development area over western Ireland with another centred about 57°N. 22°W. The depression did deepen a further 8 mb. and move to Ireland in the ensuing 12 hours, and there was a slight fall of pressure in the vicinity of the other cyclonic development pattern over the 24 hour period centred at 0300 G.M.T. on the 4th. The existence of a very strong anticyclonic development pattern further south in the Atlantic might well have damped down the cyclonic development about 57°N. 22°W. and was itself connected with marked rises of pressure.

Other features of Fig. 16b are much less intense than those already mentioned apart from one large negative value over north-eastern Italy which was only connected with small falls of surface pressure. The cyclonic development area over the English Channel would indicate that rainfall from the occlusion already over Great Britain was likely to be heaviest in this region, but the approach of the next frontal system was so rapid that it is difficult to obtain much evidence on this point.

/At

At 1500 G.M.T. on the 4th the main cyclonic development area was computed to be due north of the depression with a tongue extending into the Irish Sea. The depression actually moved NE into the north Irish Sea. There was another area of cyclonic development over the North Sea culminating in a maximum in the Straits of Dover. This was associated with the main frontal trough.

The anticyclonic development pattern behind the front is associated with a rise of pressure. The section of the Atlantic west of Ireland between 50° and 60°N. has a mainly anticyclonic development pattern and in fact a widespread pressure rise occurred in this area but there is no surface counterpart to the cyclonic development pattern about 52°N. 20°W.

The large positive value of small area about 44°N. 17°W. is not what one would expect from geometrical consideration of the contour and thickness patterns in the vicinity but has been thoroughly checked by hand computation. It emphasises the disadvantages of using interpolated contour heights and thicknesses at grid points when strong gradients are involved, but in the 12 charts computed there seem to be very few spurious values obtained from this cause.

0300 G.M.T. 3rd July, 1952 to 0300 G.M.T. 4th July, 1952.

This period was chosen for examination because of the unpredicted prolonged outbreak of rain which occurred over S.E. England and the London area on July 3rd, when pressure was markedly rising over this region. The relevant synoptic charts are shown in Figs. 19a, 20a, 21a, 22a and 23a. The development charts computed for this period are shown in Figs. 20b, 21b and 22b.

Fig. 20b shows a weak tongue of cyclonic development extending from a maximum over France across southern England and Wales at 0300 G.M.T. on the 3rd. This is still in existence 12 hours later, even though at this time the pressure was rising over the area concerned and it would appear that this is a case where upward vertical motion is associated with a surface isallobaric low but nevertheless rising surface pressure tendencies.

A misleading feature of the development chart for 0300 G.M.T. on the 4th is the increase in the cyclonic development over southern England, as the rain in this area had then ceased. Thus development charts for 0300 G.M.T. and 1500 G.M.T. on the 3rd would probably have made the forecaster suspicious about forecasting dry weather for S.E. England, but the forecaster confronted with the development chart for 0300 G.M.T. on the 4th could hardly fail to be seriously misled as regards rainfall over southern England, although there was some increase in the surface troughing over S.W. England.

The other features of this sequence of development charts agree fairly well with actuality although there are some small computed features which do not appear to be connected with any actual atmospheric development.

General Results.

All development areas numerically greater than 40 units have been classified according to their degree of usefulness to a forecaster, and the results are shown in the appendix. Each of these development areas has been placed in one of the following classes:-

A = useful guidance.

B = not particularly useful, but not misleading.

C = misleading guidance.

It should be noted that the B classification has been used mainly for those development areas which were greater than 40 units, but which were overshadowed by other development areas on the same chart. Although this is a very subjective classification, it gives some idea of the usefulness of development charts, and a summary of the results is shown in Table I.

/Table I

Table I.
No. of cases in each class.

A	B	C
41	36	13

There are many more development areas in class A than C, whilst most of those in class B would be neglected by forecasters with experience in the use of development charts.

Conclusion.

Although only 12 development charts have been computed there are certain conclusions which one can come to about the technique involved. The main conclusions are enumerated below:-

1. It is now possible to compute the Sutcliffe development term on LEO for 96 points in 5½ minutes. The total time, which includes reading data from charts, punching and checking data tapes, transmitting data from A to B and the answer from B to A is estimated to be about 1 hour.

2. The development patterns computed by such a method agree quite well in shape with those computed when the Sawyer - Matthewman scale is used, although the maximum values are rather less owing to smoothing over a larger area.

3. The more intense features of the computed development term are most frequently though not always associated with an appropriate change in the synoptic situation. There is probably some connection between rainfall and maximum cyclonic development areas, but this aspect has not been stressed in this report.

4. Features of lesser intensity are difficult to assess. They may be unimportant features which have been arbitrarily brought into the computations, either as a result of difficulties of interpolation in reading data from the charts or by reading off data from apparently consistent contour and thickness charts implying surface features which do not exist. Alternatively these features may be real, in which case one has to decide whether they are important in their own right or whether they are likely to be damped out by adjacent more intense features.

Thus given the necessary computing facilities it is now possible for the forecaster to have development charts computed in a sufficiently short space of time to be of use. It seems probable that such charts would be of some use to the forecaster, and certainly of increasing use as the forecaster became experienced in interpreting them, but until it is possible to use such charts under working conditions, it is difficult to estimate how much they would improve forecasting.

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APPENDIX

A = useful guidance; B = not particularly useful, but not misleading; C = misleading guidance.

Date	Position of Development area	Classification
1500 G.M.T. 14. 3.49	53°N 32°W	B
"	62°N 16°W	A
"	53°N 10°W	A
"	59°N 5°W	A
"	62°N 4°W	B
"	53°N 2°E	A
1500 G.M.T. 19.10.50	52°N 30°W	A
"	62°N 5°W	A
"	46°N 2°W	B
"	47°N 9°E	B
0300 G.M.T. 20.10.50	54°N 30°W	A
"	62°N 7°E	A
"	59°N 18°E	A
"	50°N 12°E	B
"	46°N 12°E	B
1500 G.M.T. 20.10.50	57°N 30°W	A
"	45°N 26°W	C
"	43°N 25°W	B
"	62°N 8°E	A
0300 G.M.T. 21.10.50	60°N 30°W	A
"	54°N 32°W	A
"	45°N 27°W	A
"	60°N 12°E	A
"	58°N 21°E	C
2100 G.M.T. 14. 2.50	52°N 30°W	C
"	48°N 28°W	C
"	56°N 19°W	B
"	57°N 10°W	A
"	45°N 2°E	B
"	50°N 4°E	A
"	47°N 13°E	B
"	51°N 16°E	B
0300 G.M.T. 15. 2.50	54°N 27°W	B
"	47°N 14°W	B
"	56°N 19°W	B
"	60°N 15°W	A
"	57°N 1°E	A
"	56°N 5°E	B
"	50°N 8°E	A
"	45°N 7°E	A
0300 G.M.T. 4. 2.51	47°N 28°W	B
"	52°N 23°W	A
"	56°N 22°W	B
"	48°N 21°W	B
"	46°N 20°W	B

Date	Position of Development area	Classification
0300 G.M.T. 4. 2.51	54°N 10°W	A
"	57°N 10°W	B
"	50°N 2°W	A
"	47°N 1°E	B
"	43°N 9°E	C
"	45°N 10°E	C
1500 G.M.T. 4. 2.51	50°N 30°W	B
"	42°N 27°W	A
"	50°N 24°W	A
"	44°N 22°W	C
"	52°N 19°W	B
"	44°N 16°W	B
"	52°N 13°W	A
"	57°N 9°W	A
"	44°N 6°W	A
"	52°N 2°W	A
"	52°N 3°E	A
"	53°N 9°E	C
"	52°N 16°E	B
0300 G.M.T. 3. 7.52	60°N 30°W	B
"	44°N 20°W	B
"	44°N 14°W	B
"	44°N 10°W	B
"	46°N 10°W	B
"	49°N 6°W	B
"	47°N 2°E	A
"	59°N 4°E	B
"	52°N 9°E	A
"	58°N 11°E	A
"	57°N 19°E	A
1500 G.M.T. 3. 7.52	58°N 33°W	A
"	45°N 15°W	B
"	45°N 3°W	C
"	50°N 1°E	A
"	54°N 4°E	B
"	44°N 11°E	A
"	50°N 13°E	C
"	58°N 15°E	A
0300 G.M.T. 4. 7.52	56°N 33°W	A
"	45°N 16°W	C
"	52°N 4°W	C
"	44°N 1°E	C
"	46°N 5°E	B
"	54°N 12°E	A
"	57°N 20°E	B

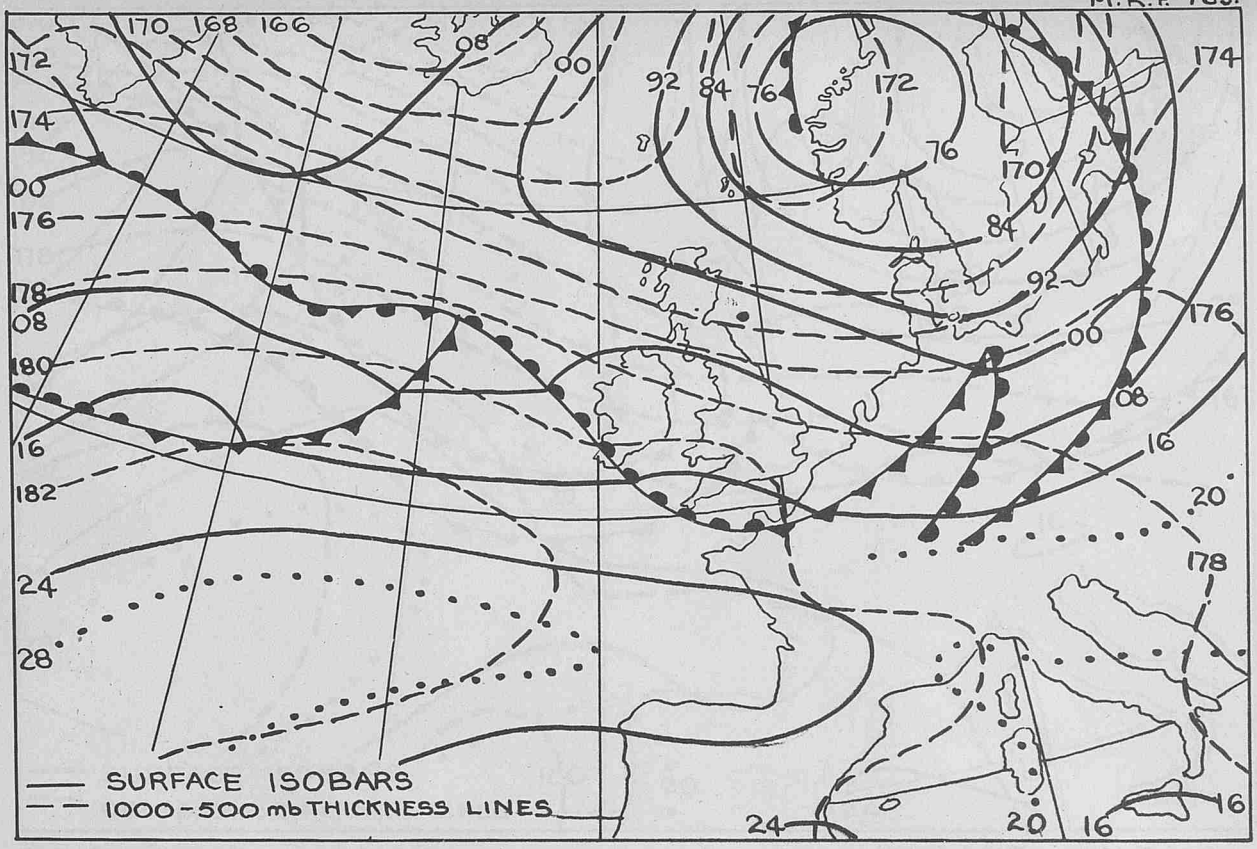


FIG 2a SYNOPTIC CHART. 0300 GMT. 14. 3. 49.

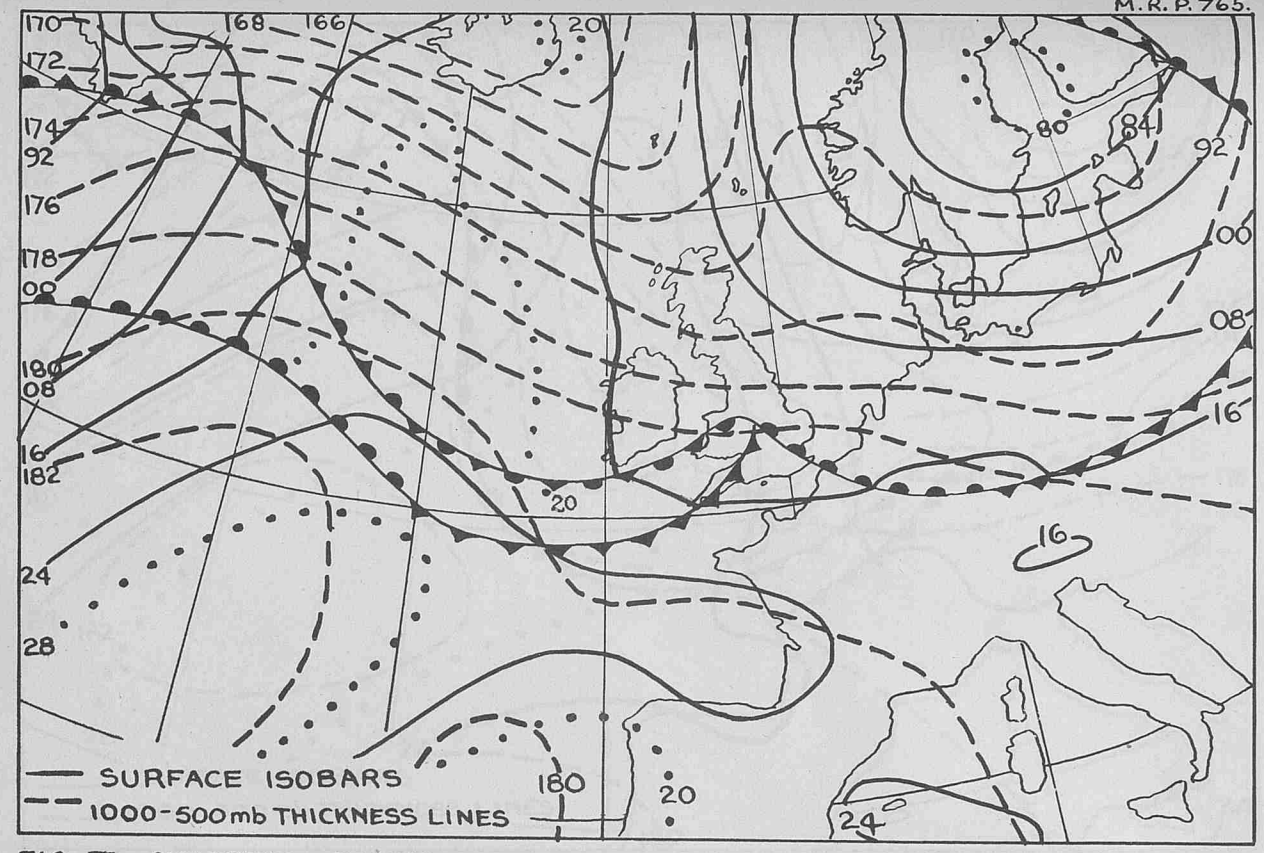


FIG. 3a SYNOPTIC CHART 1500 GMT. 14.3.49.

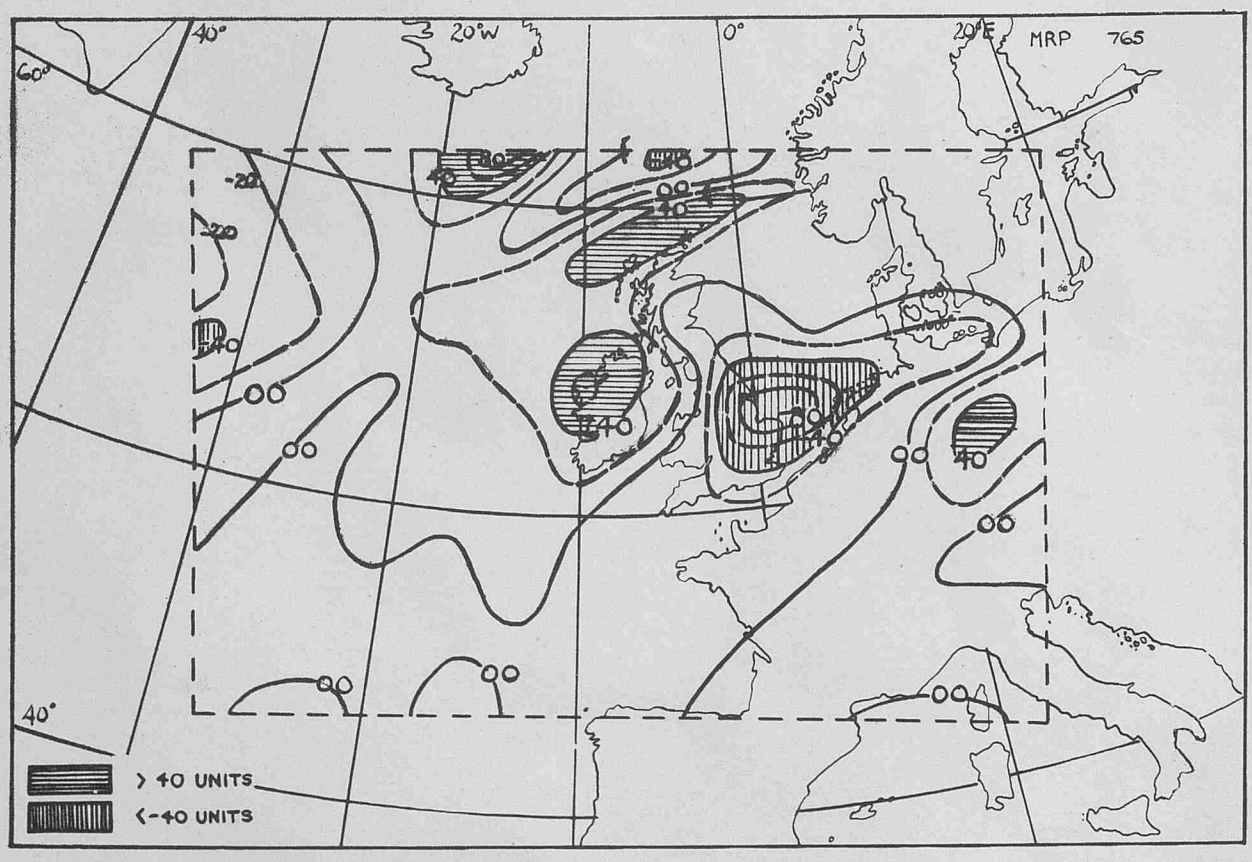


FIG. 3.b DEVELOPMENT CHART. (UNITS 10^{-3} HRS^{-2}) 1500 G M T 14·3·49

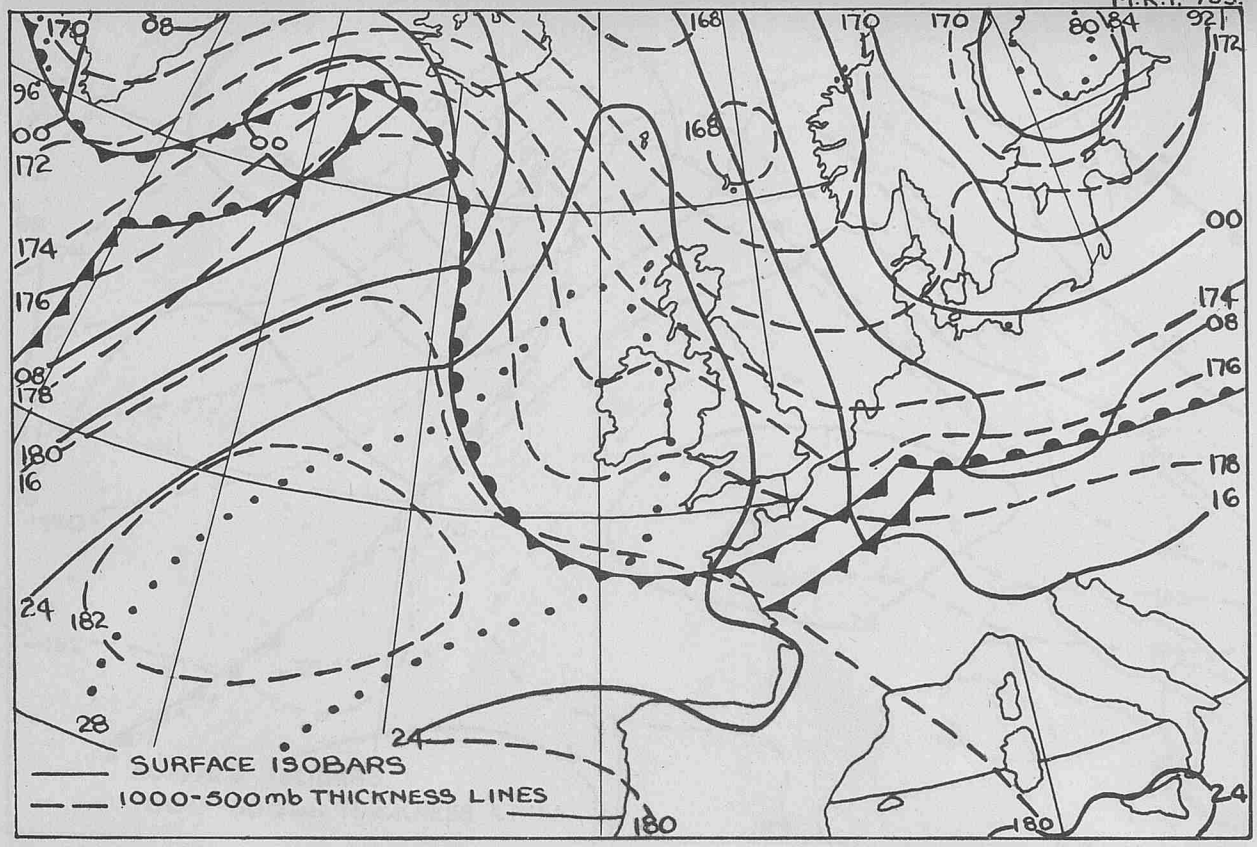


FIG. 4a SYNOPTIC CHART 0300 GMT. 15. 3. 49.

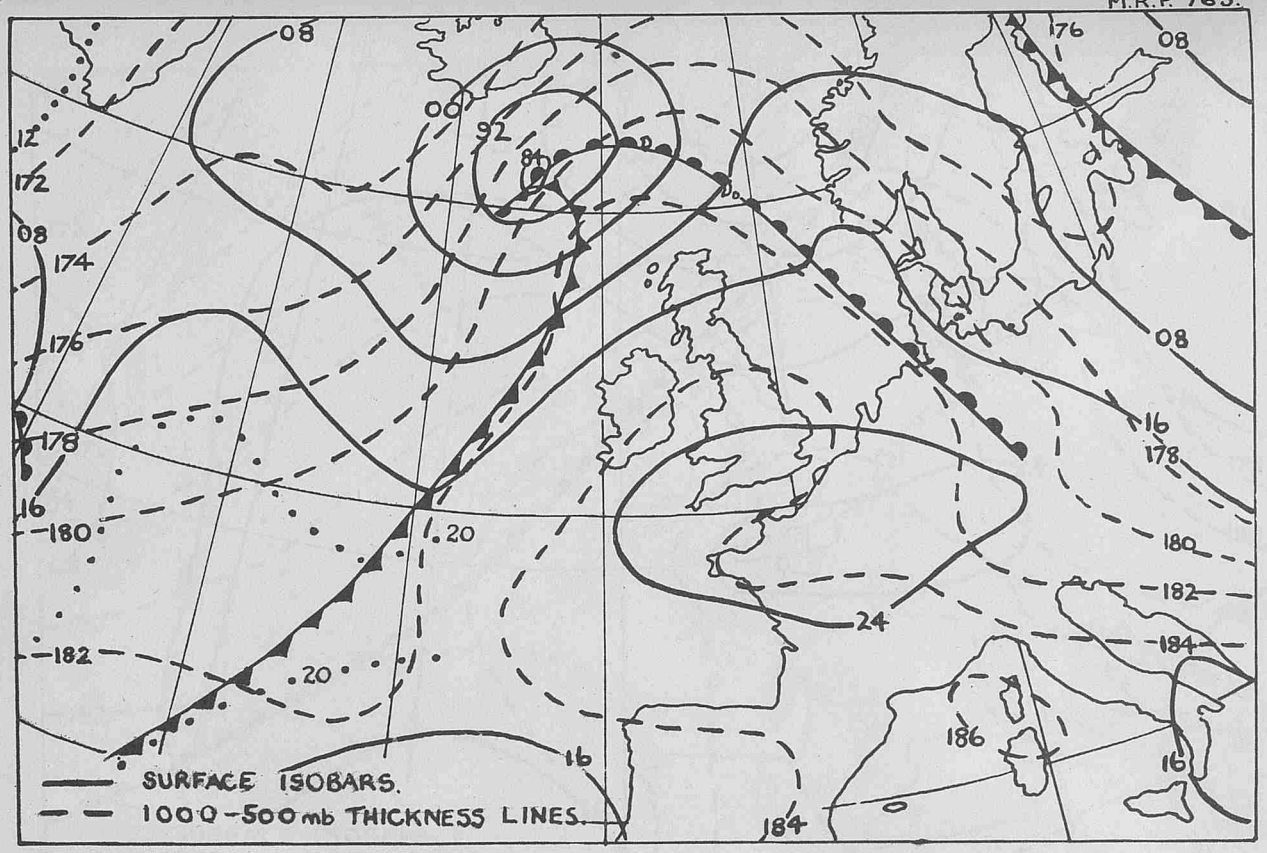


FIG. 5a SYNOPTIC CHART. 0300 GMT 19.10.50.

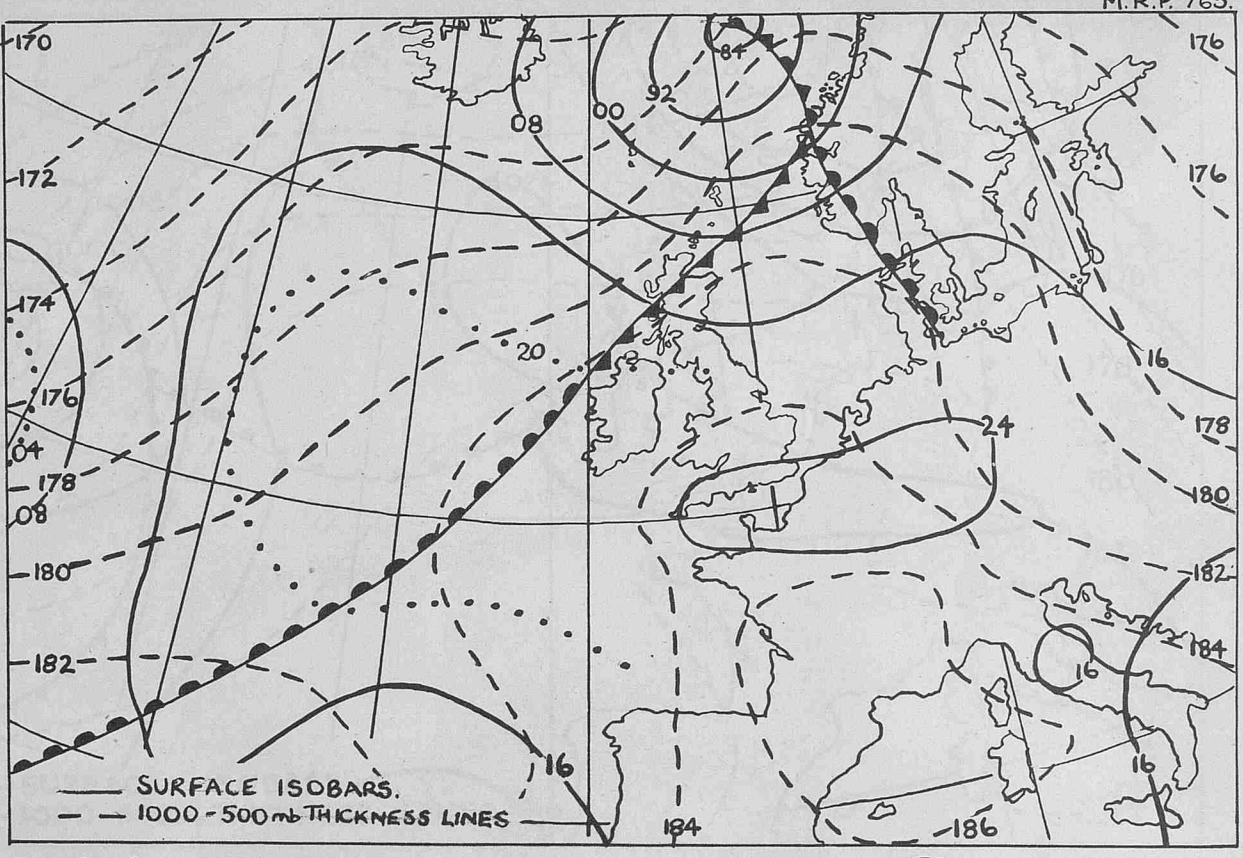


FIG. 6a SYNOPTIC CHART. 1500 G.M.T. 19.10.50.

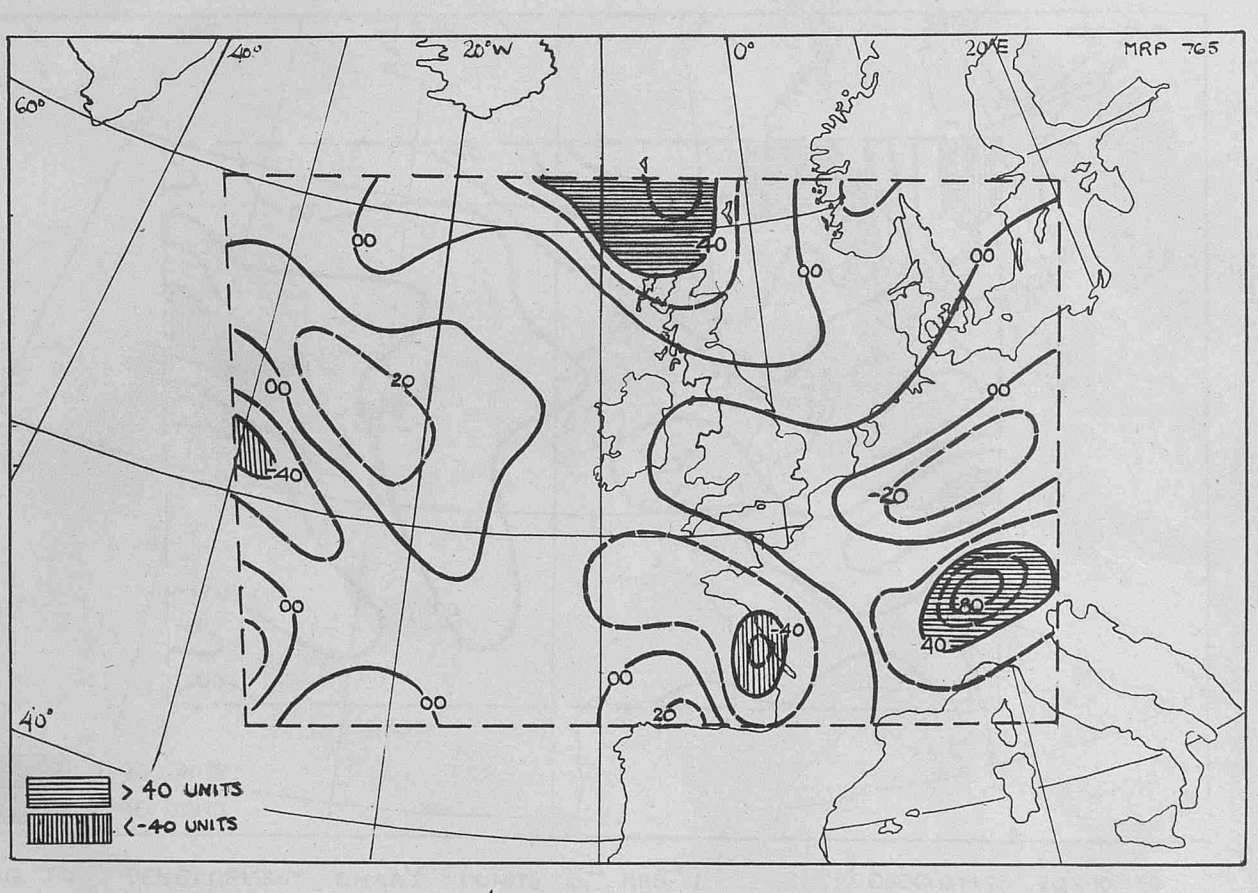


FIG 6b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 1500 G M T 19·10·50

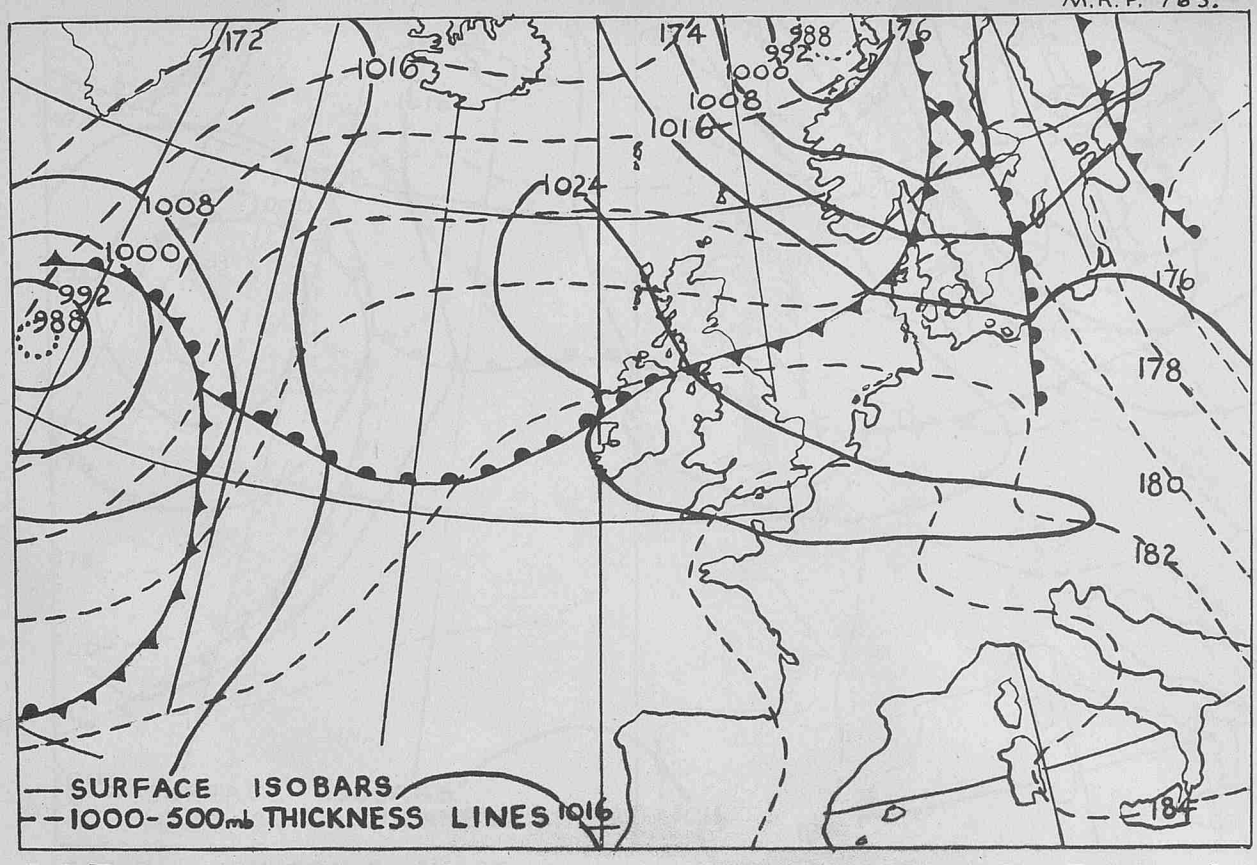


FIG 7a. SYNOPTIC CHART 0300 G.M.T. 20.10.50.

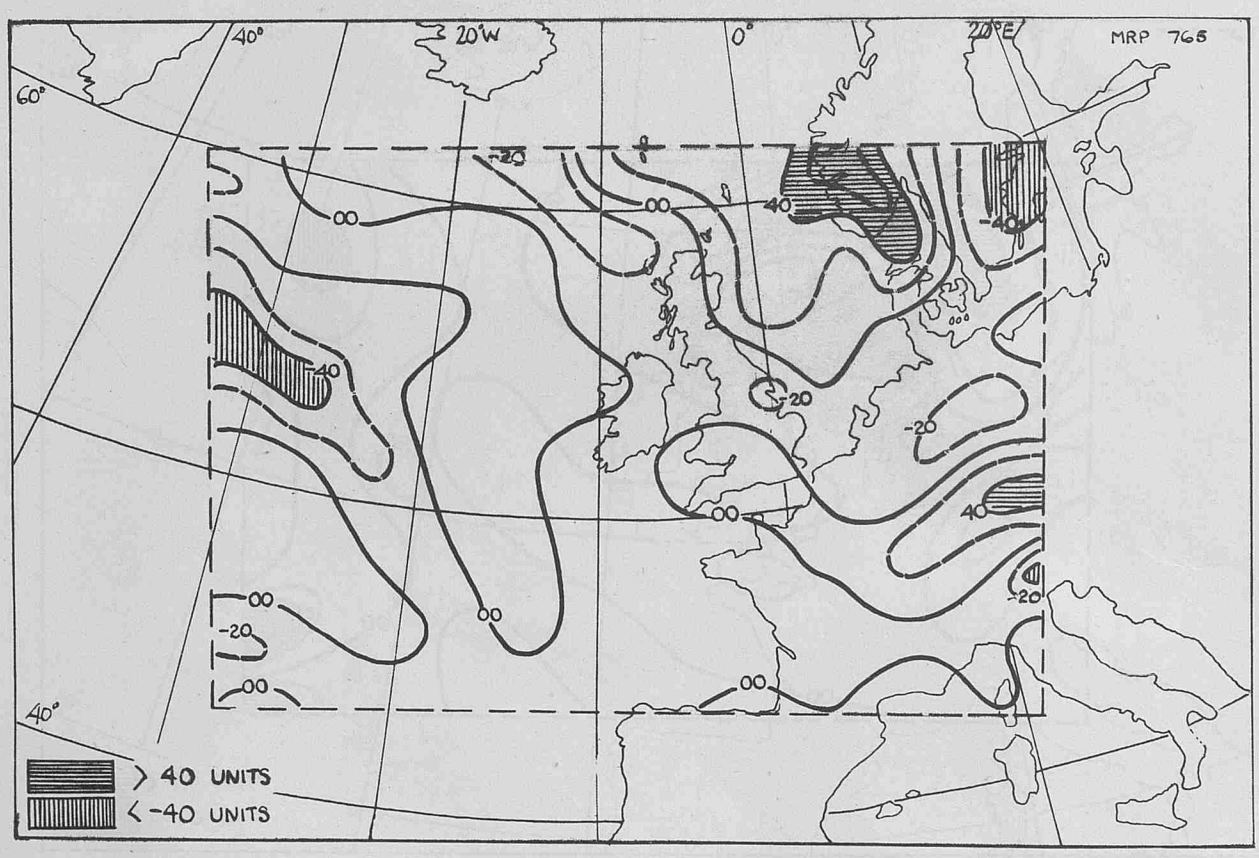


FIG. 7b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 0300 GMT 20·10·50

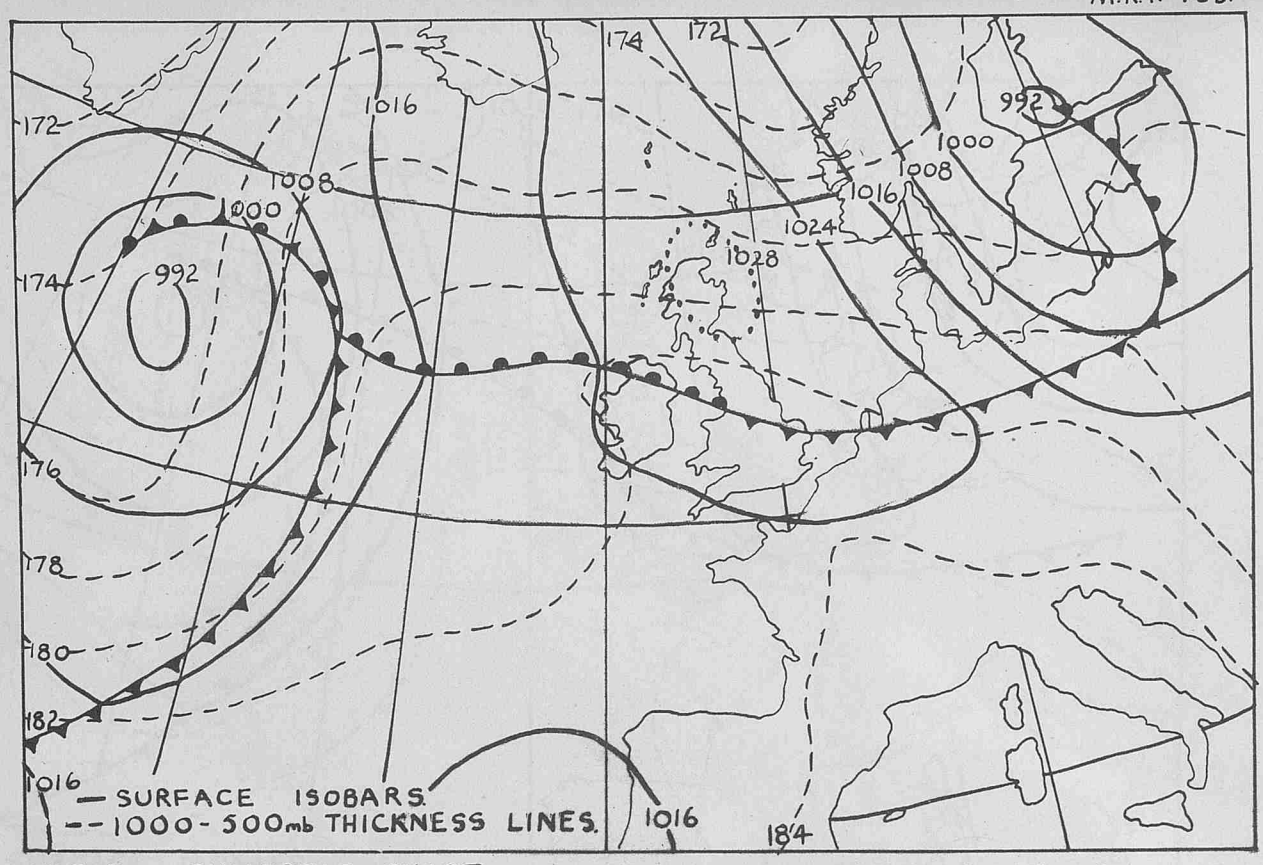


FIG. 8a SYNOPTIC CHART. 1500 GMT. 20.10.50.

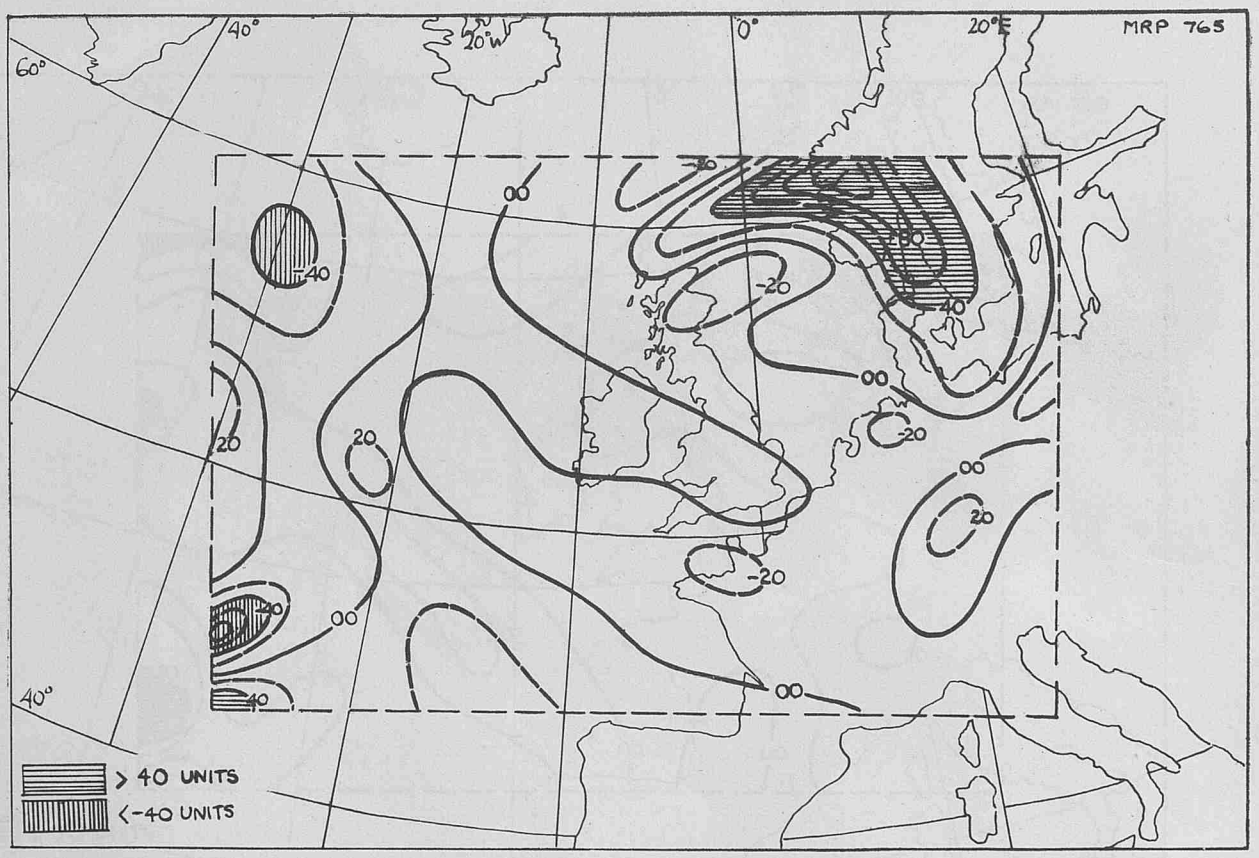


FIG 8b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 1500 GMT 20·10·50.

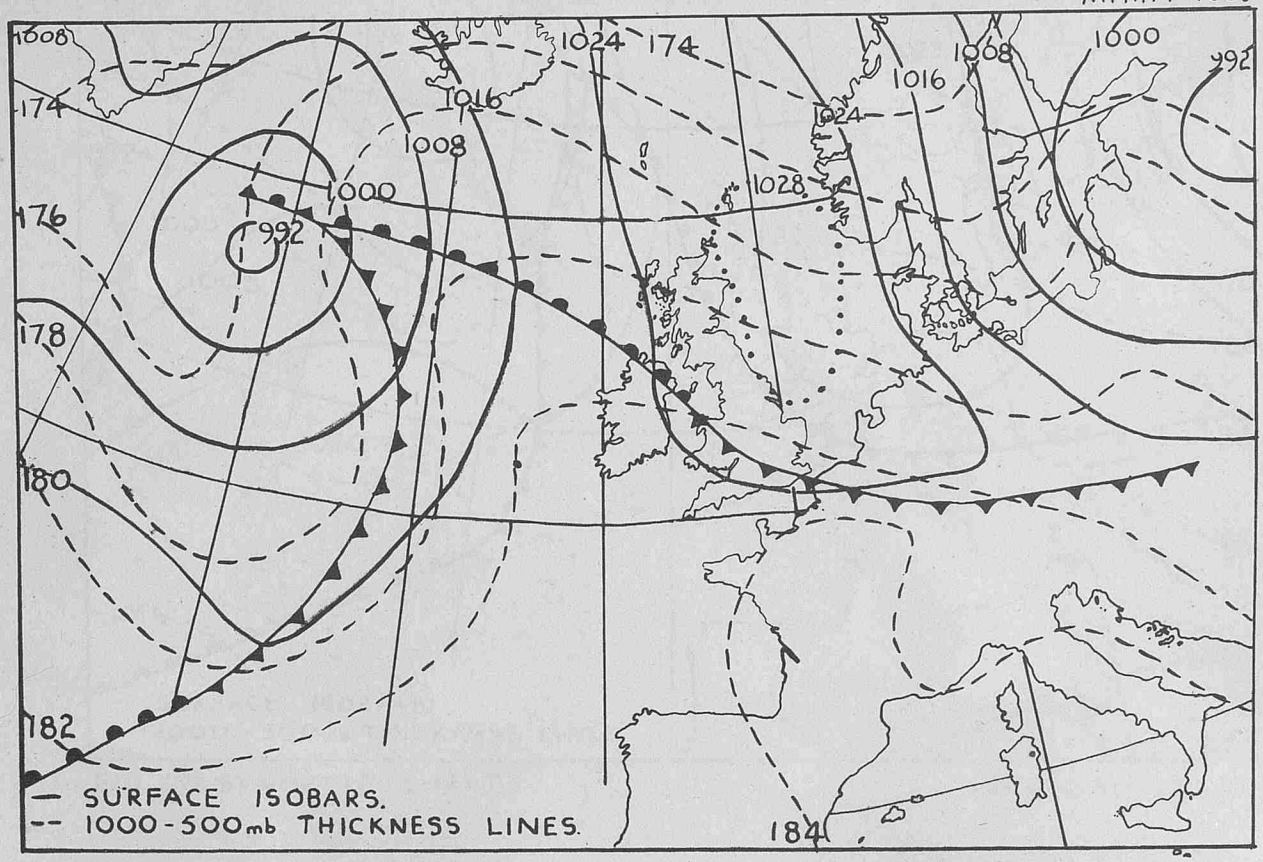


FIG. 9a SYNOPTIC CHART. 0300 GMT. 21.10.50.

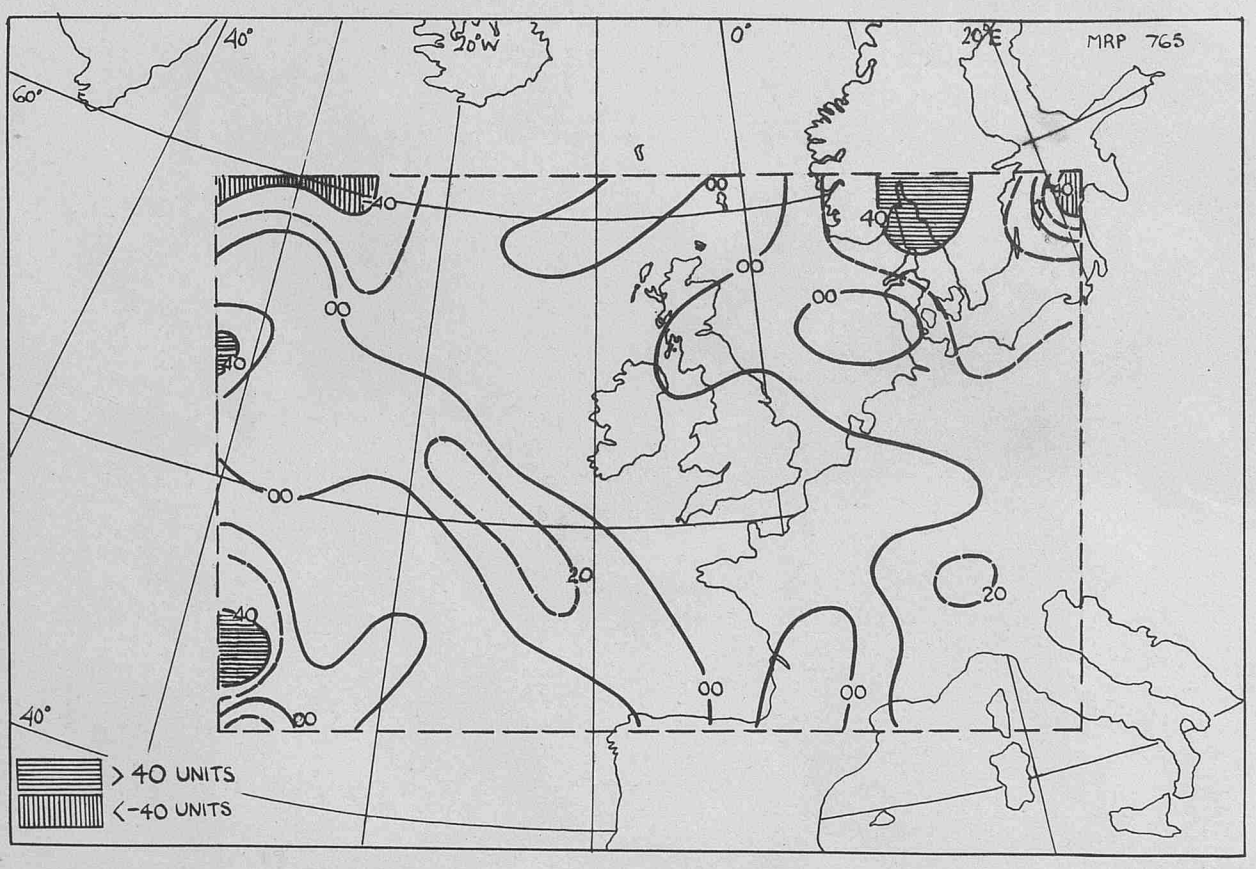


FIG 9b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 0300 GMT 21·10·50.

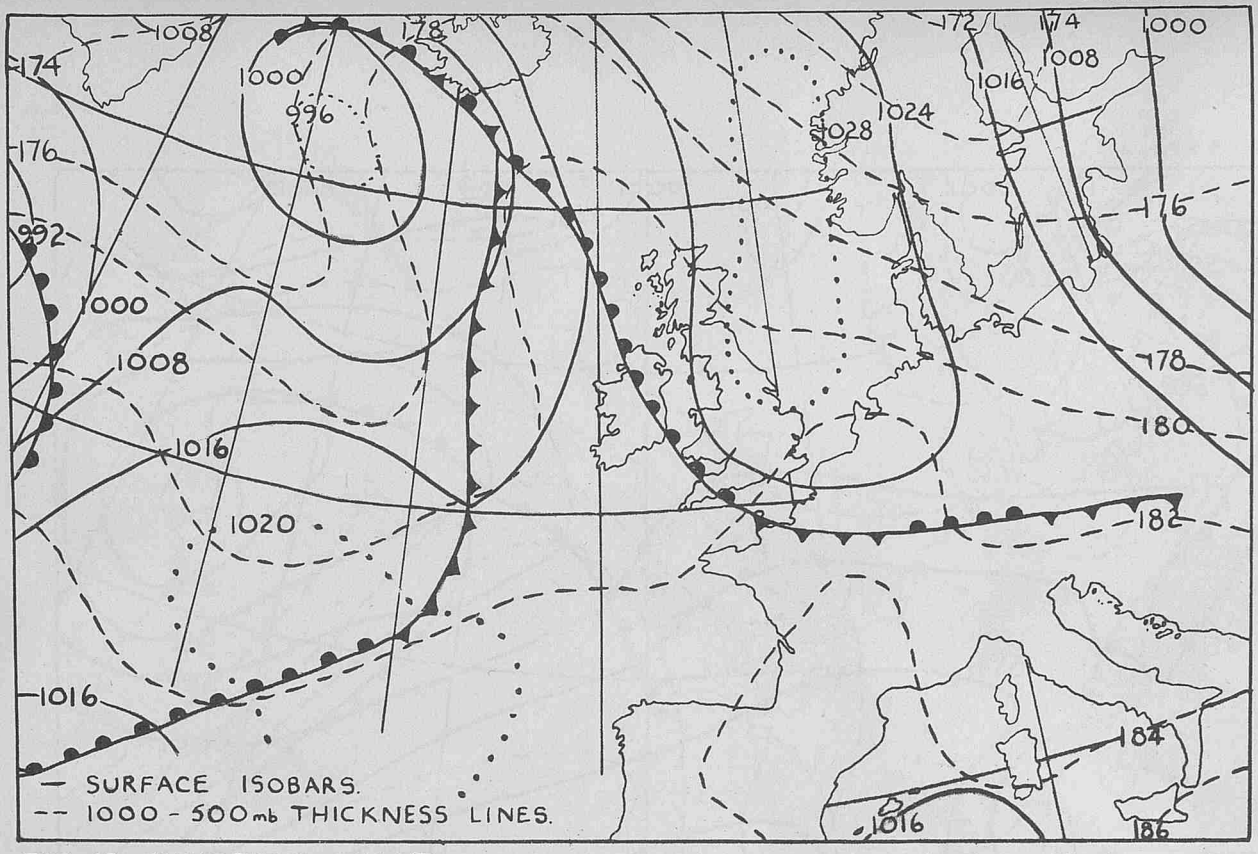


FIG. 10a SYNOPTIC CHART. 1500 GMT. 21.10.50.

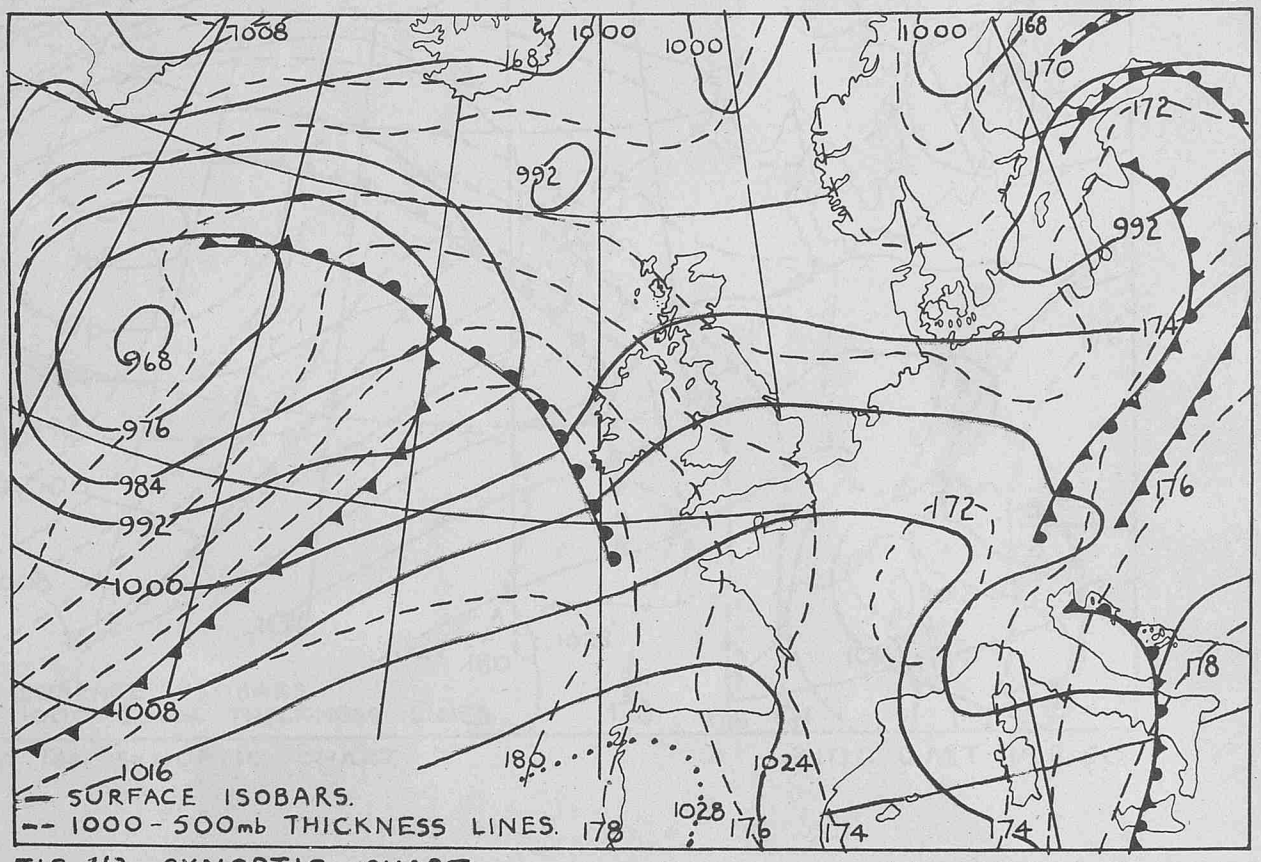


FIG 11a SYNOPTIC CHART 1500 GMT. 14.2.50.

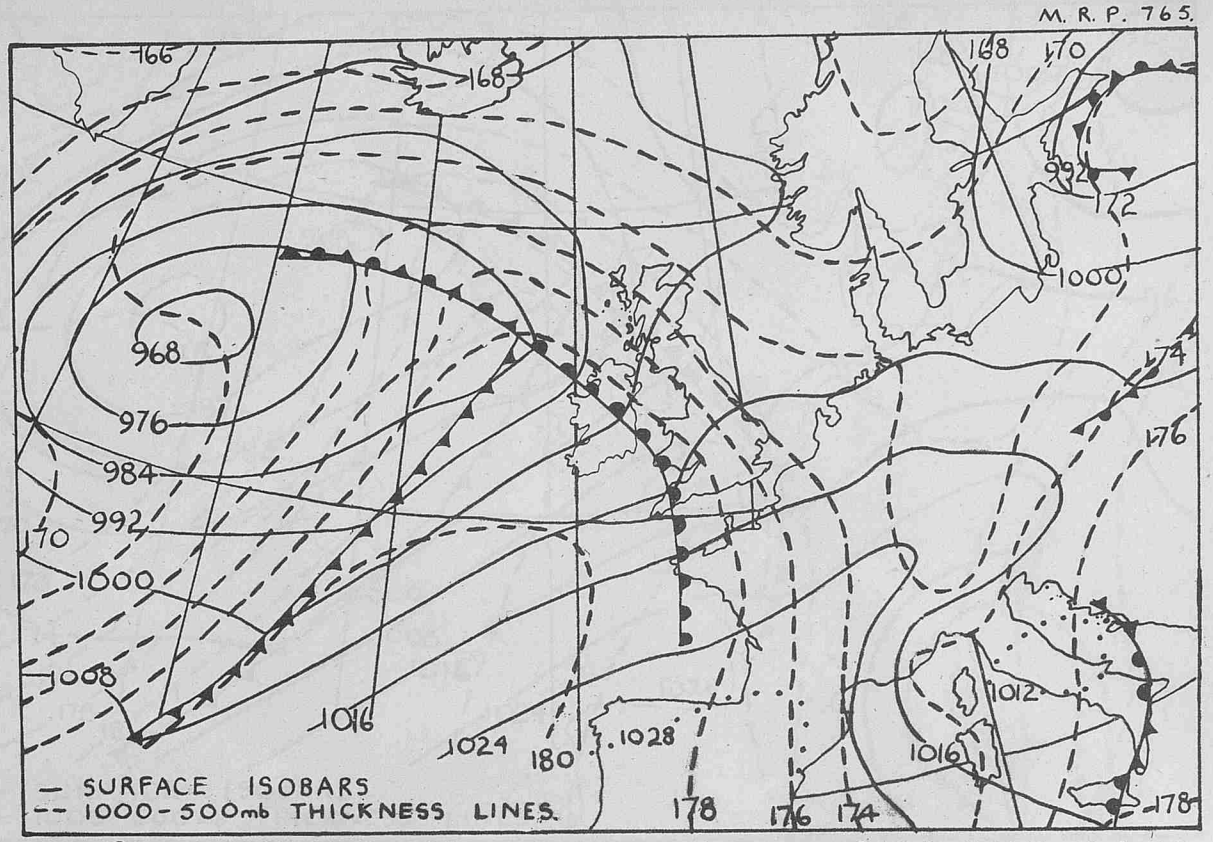


FIG 12a SYNOPTIC CHART. 2100 GMT. 14.2.50.

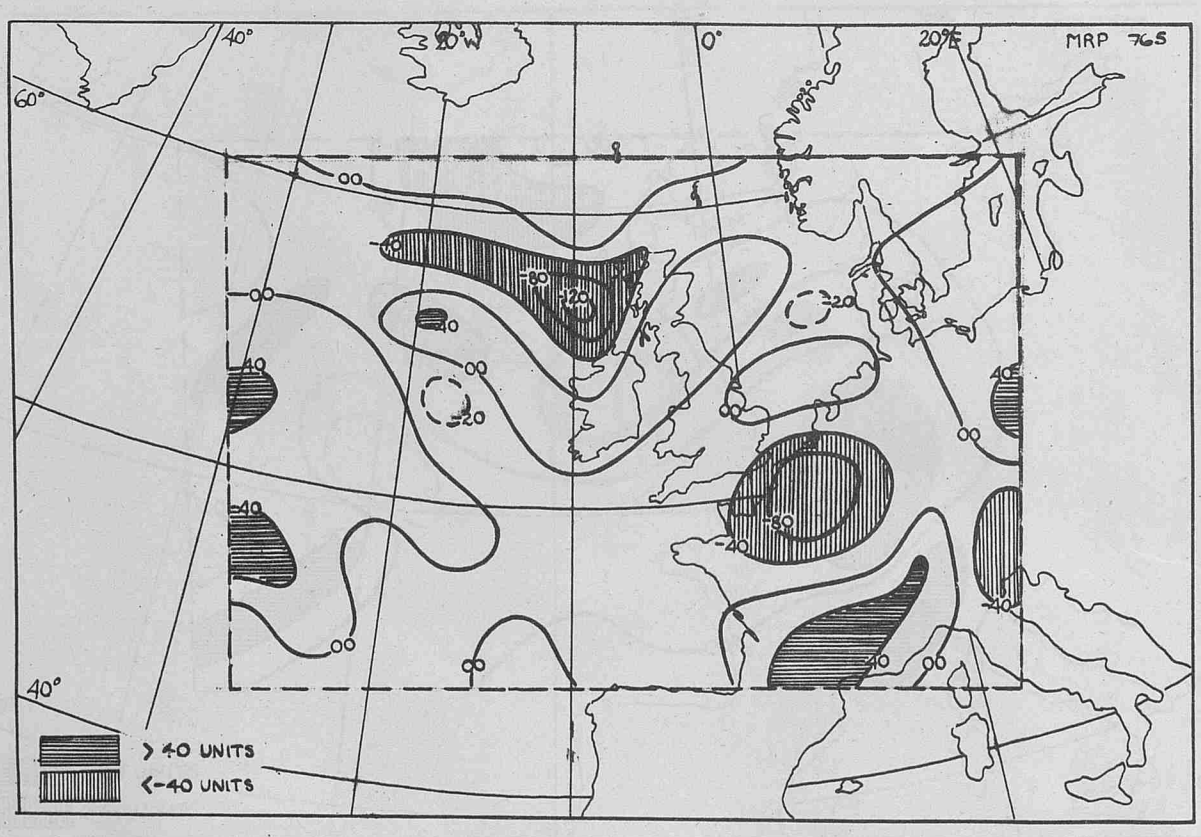


FIG. 12b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 2100 GMT 14·2·50

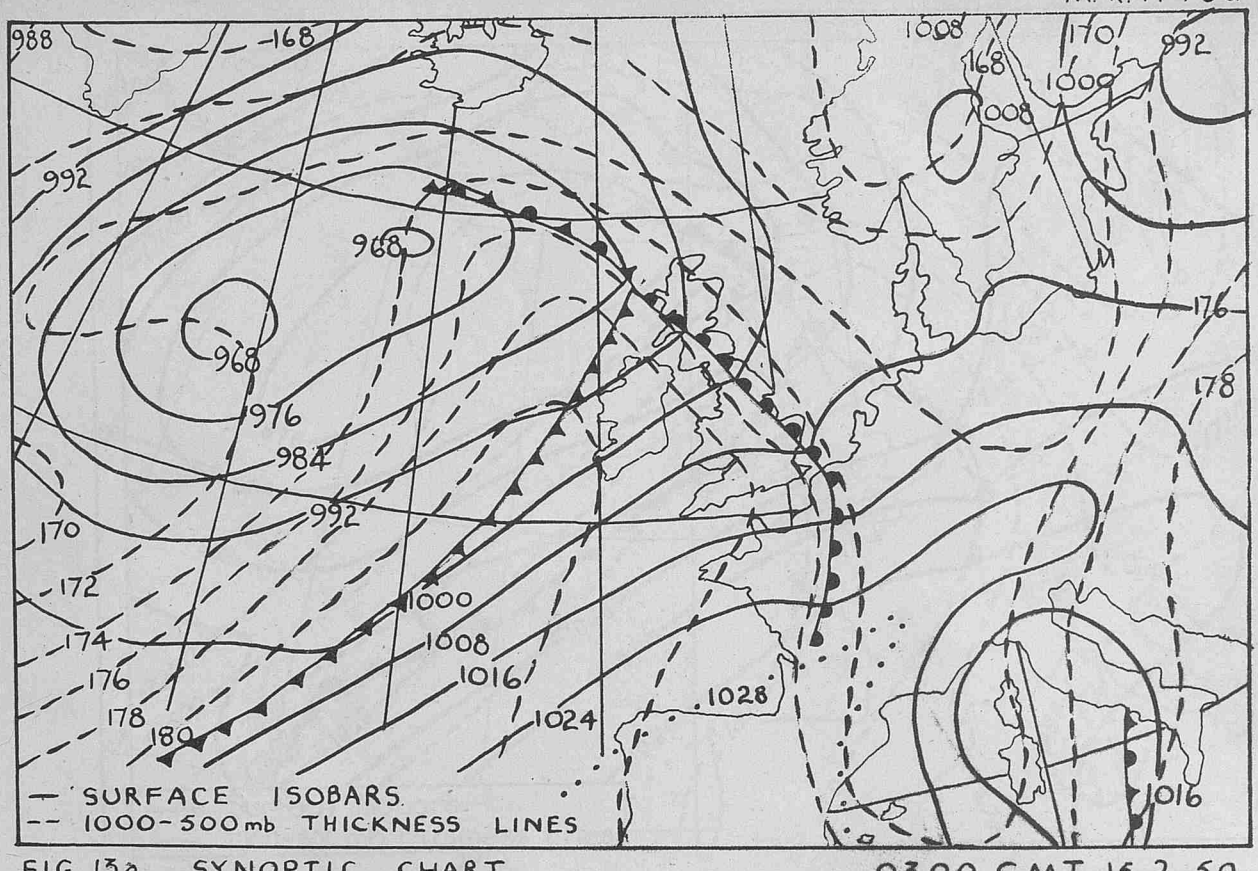


FIG 13a. SYNOPTIC CHART 0300 GMT. 15. 2. 50.

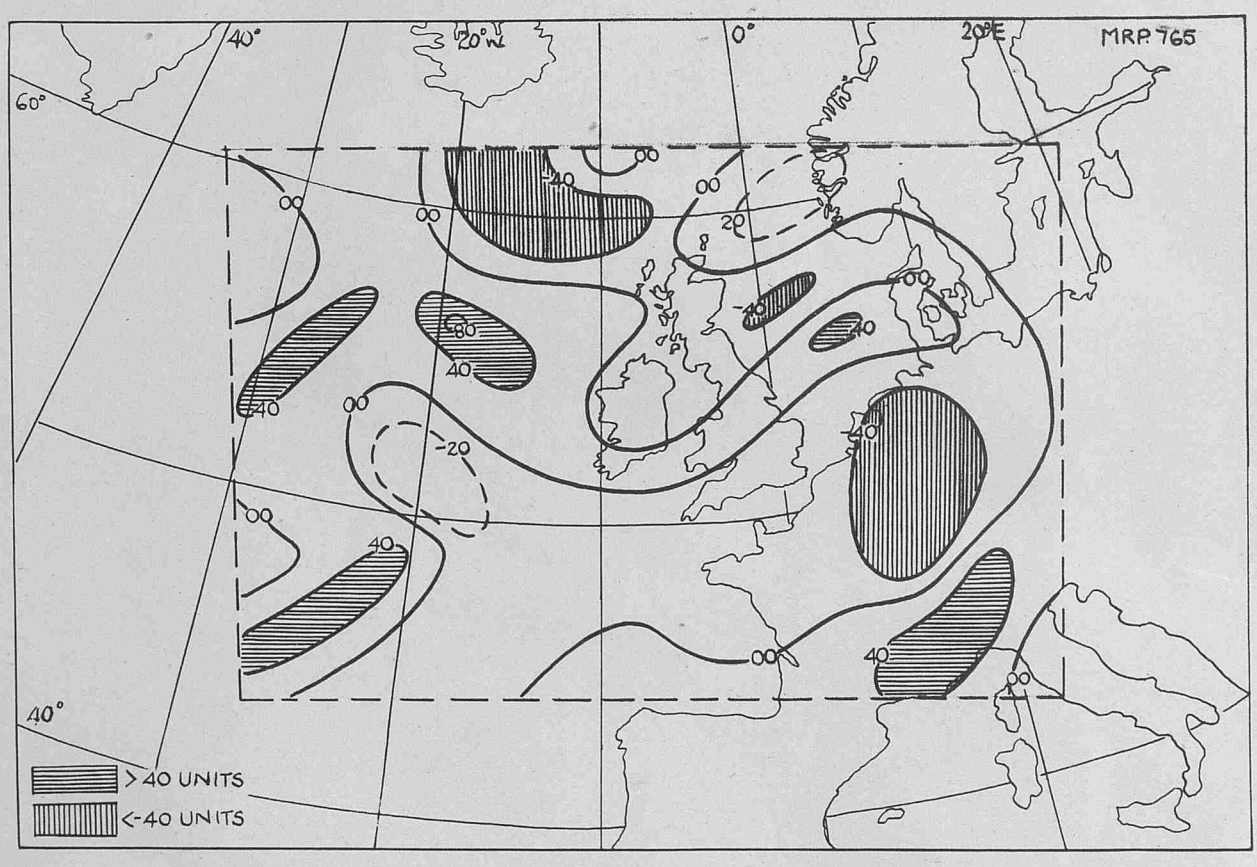
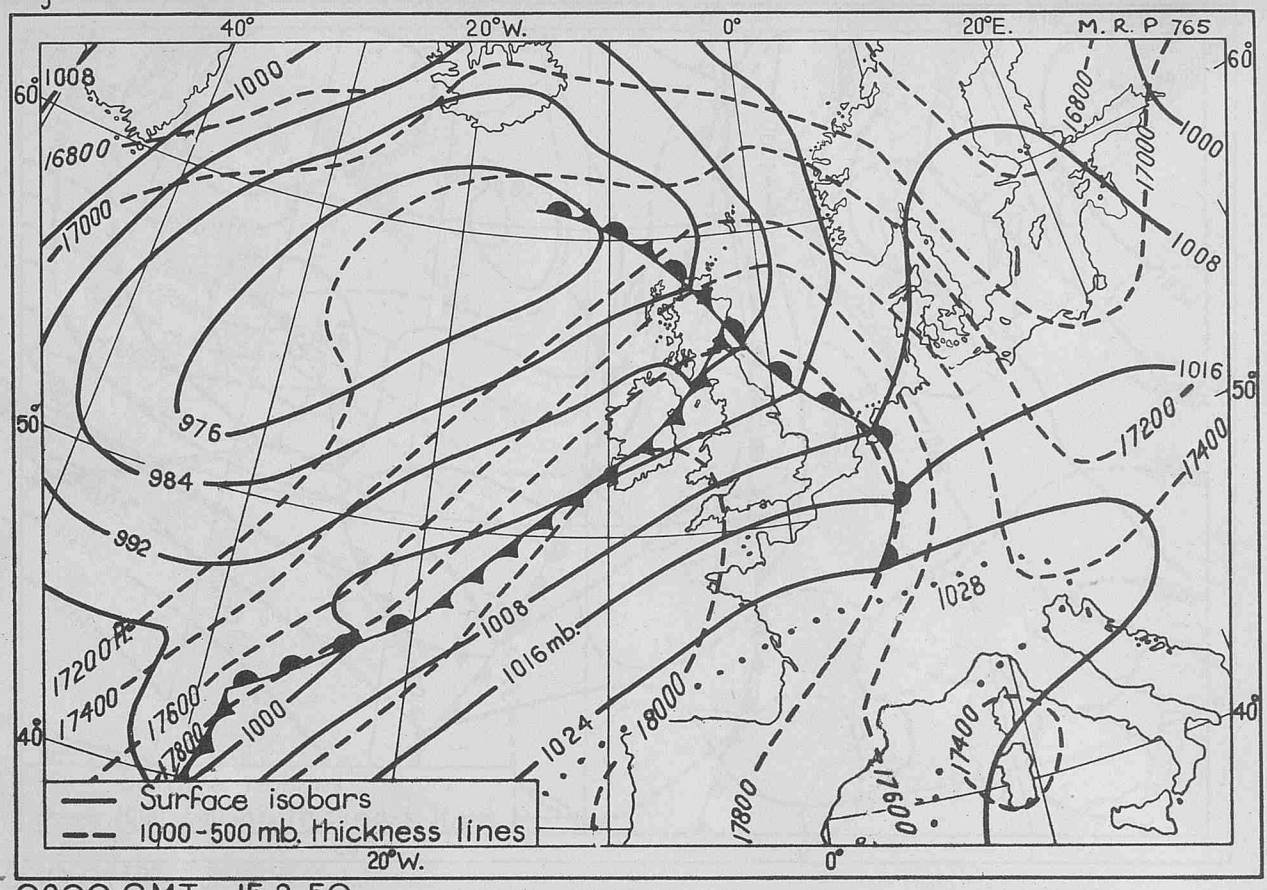


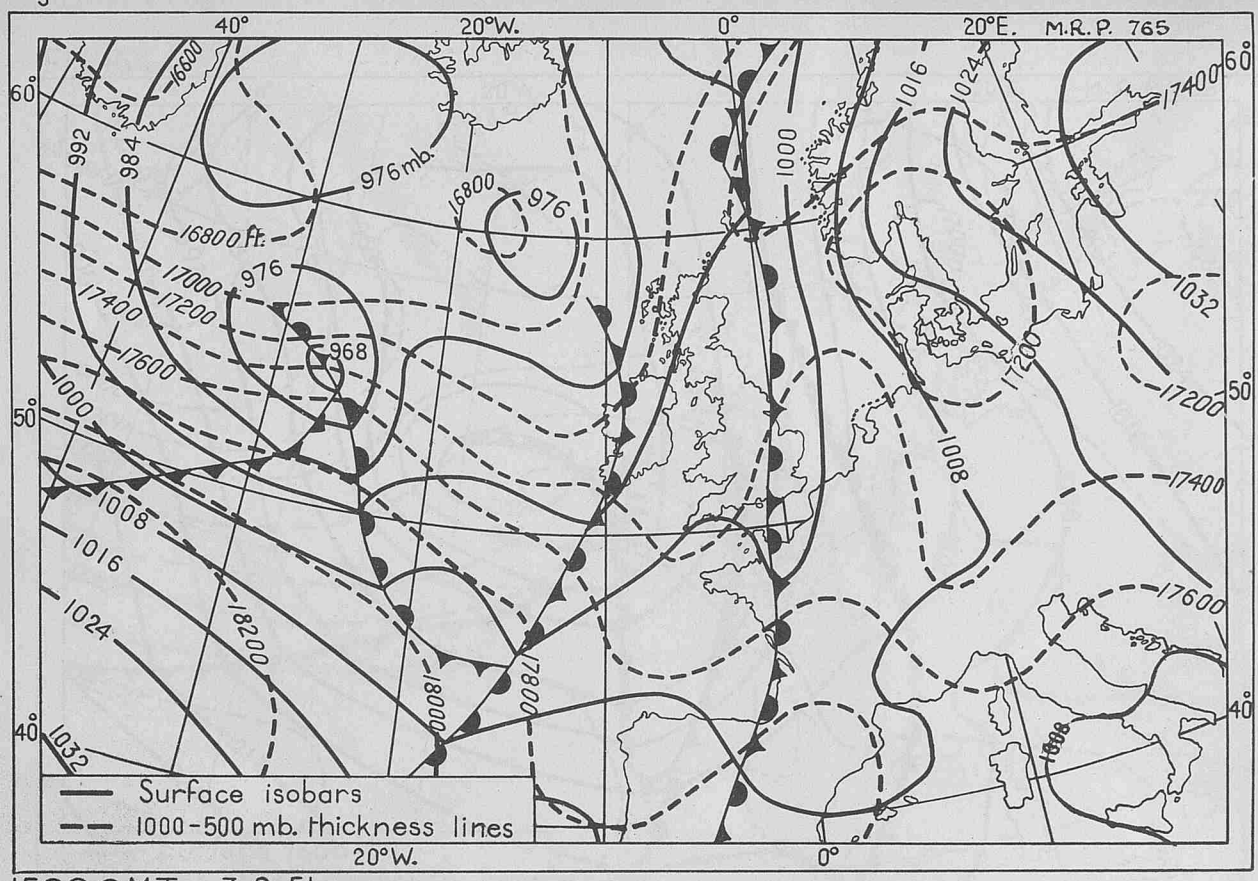
FIG 13b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 0300 GMT. 15·2·50.

Fig. 14a



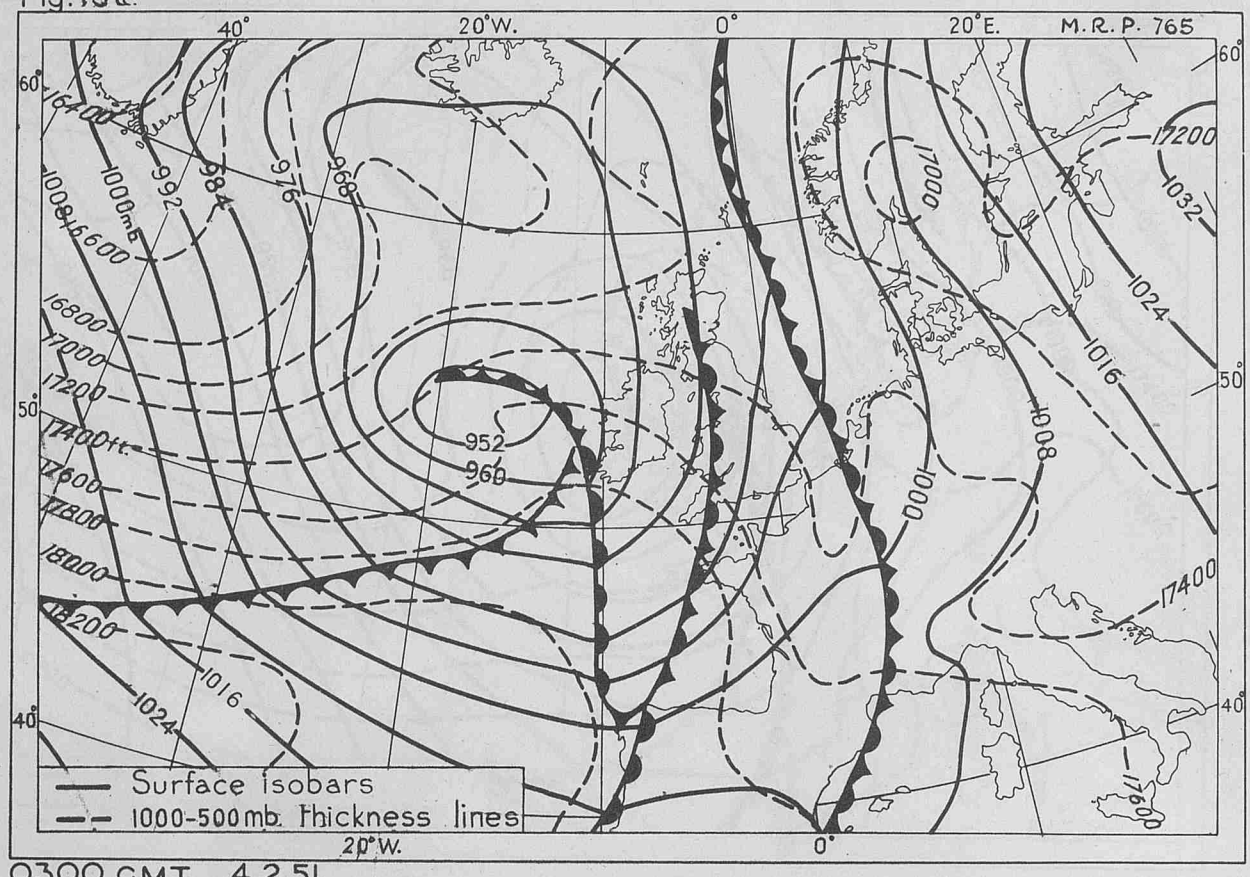
0900 G.M.T. 15.2.50

Fig. 15a



1500 G.M.T. 3.2.51

Fig. 16a.



0300 G.M.T. 4.2.51.

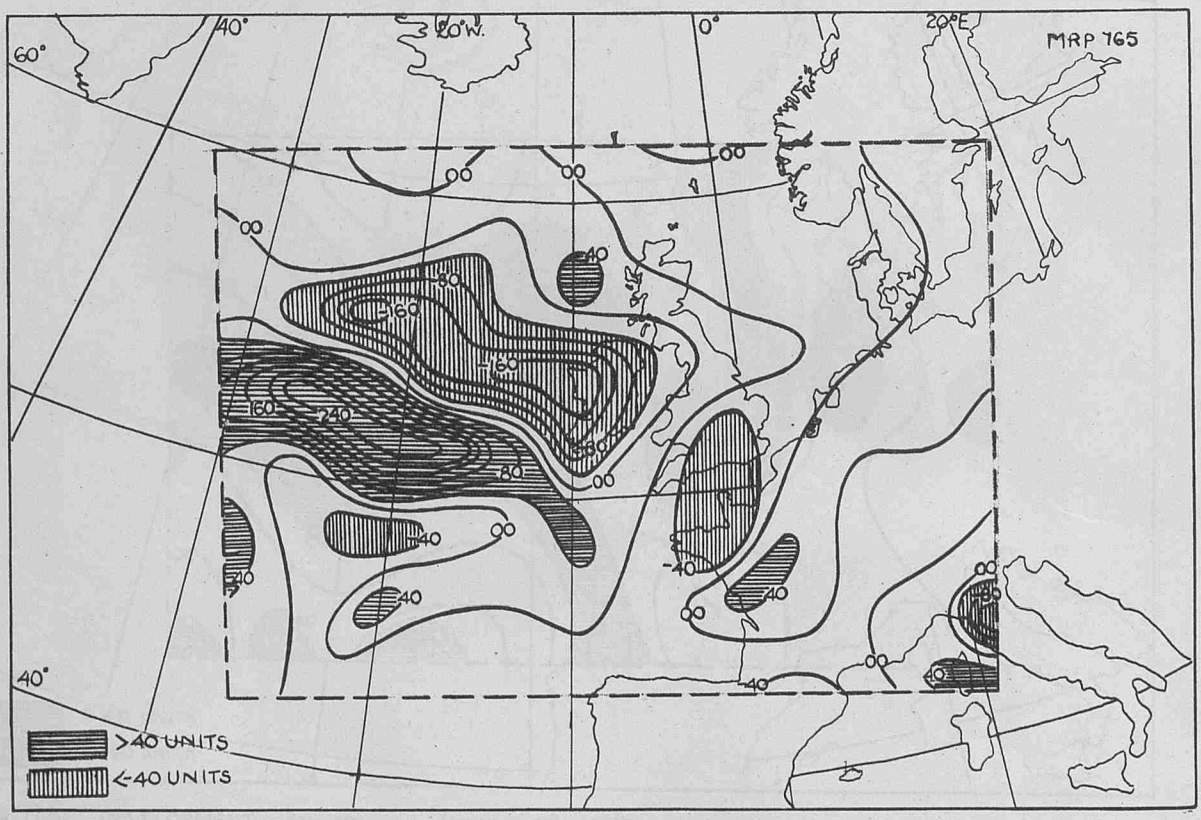
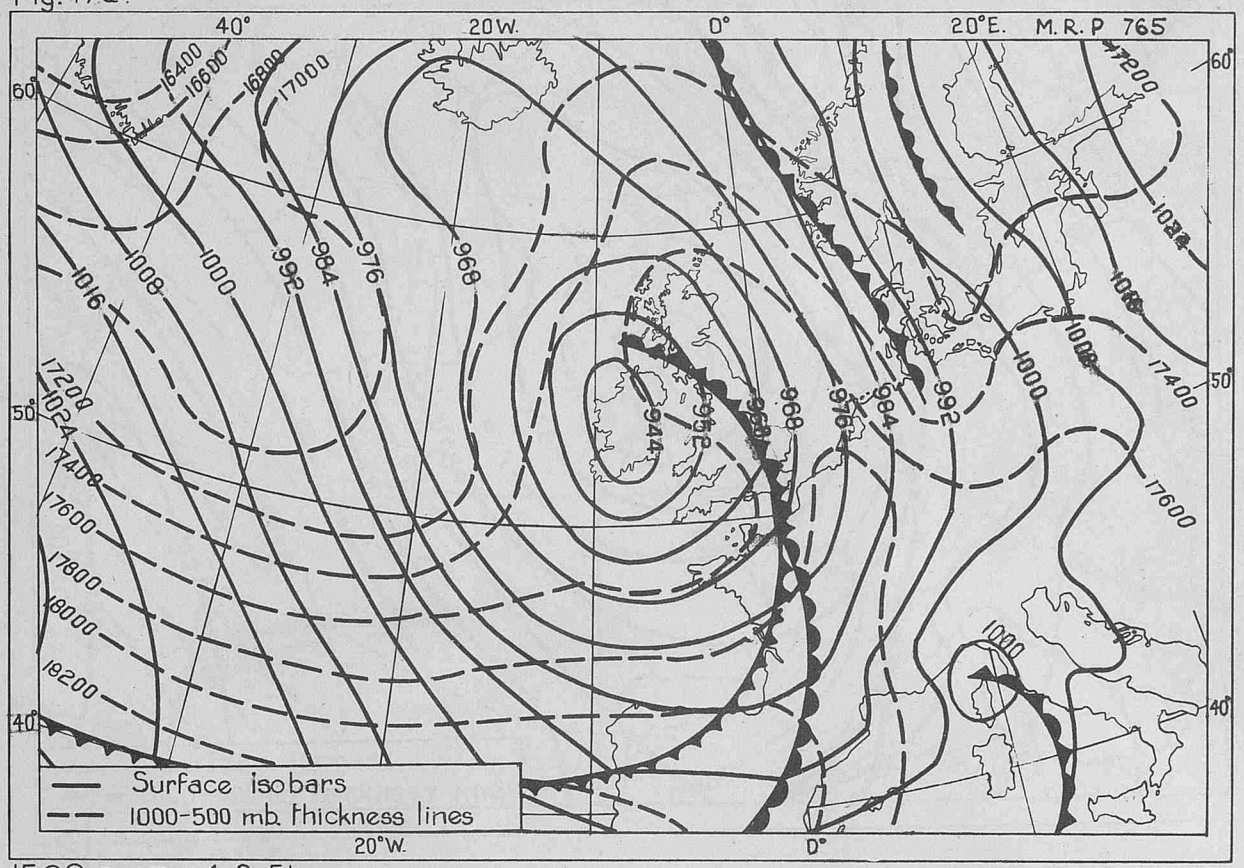


FIG 16b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 0300 GMT 4·2·51.

Fig. 17a.



1500 G.M.T. 4.2.51.

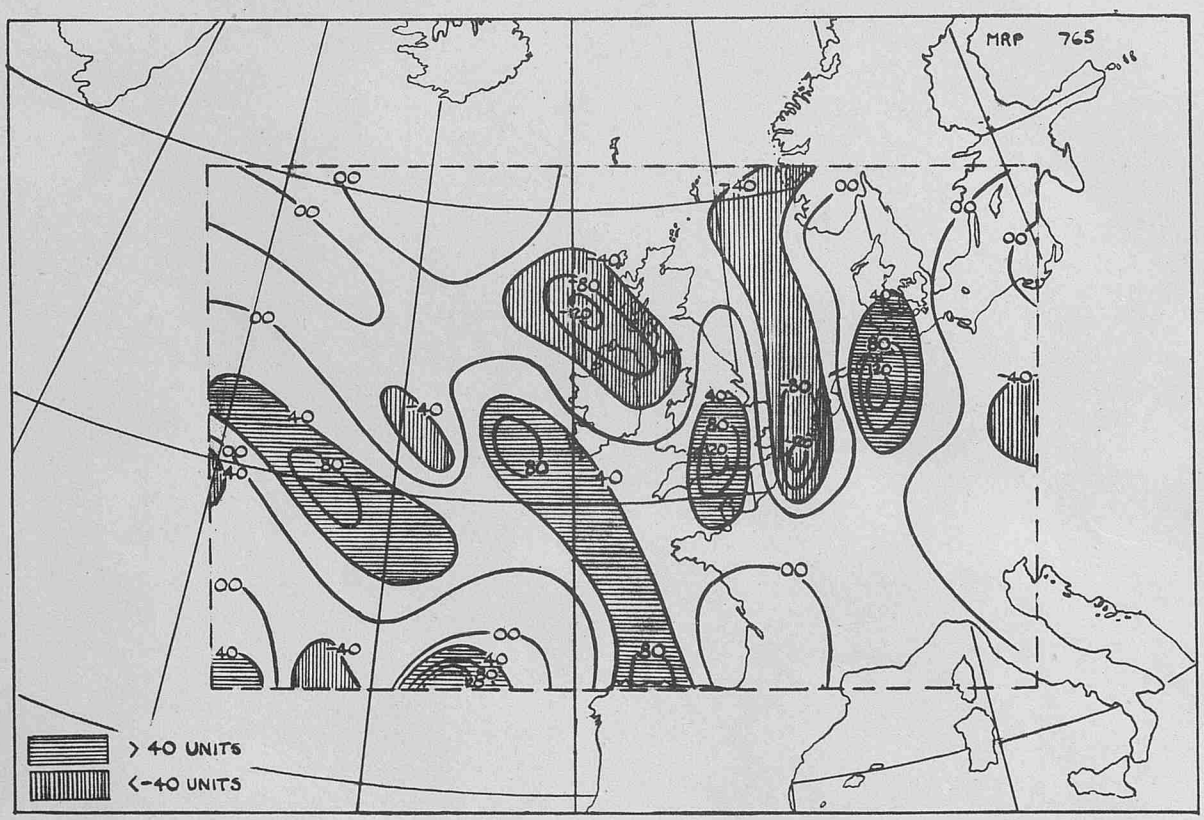
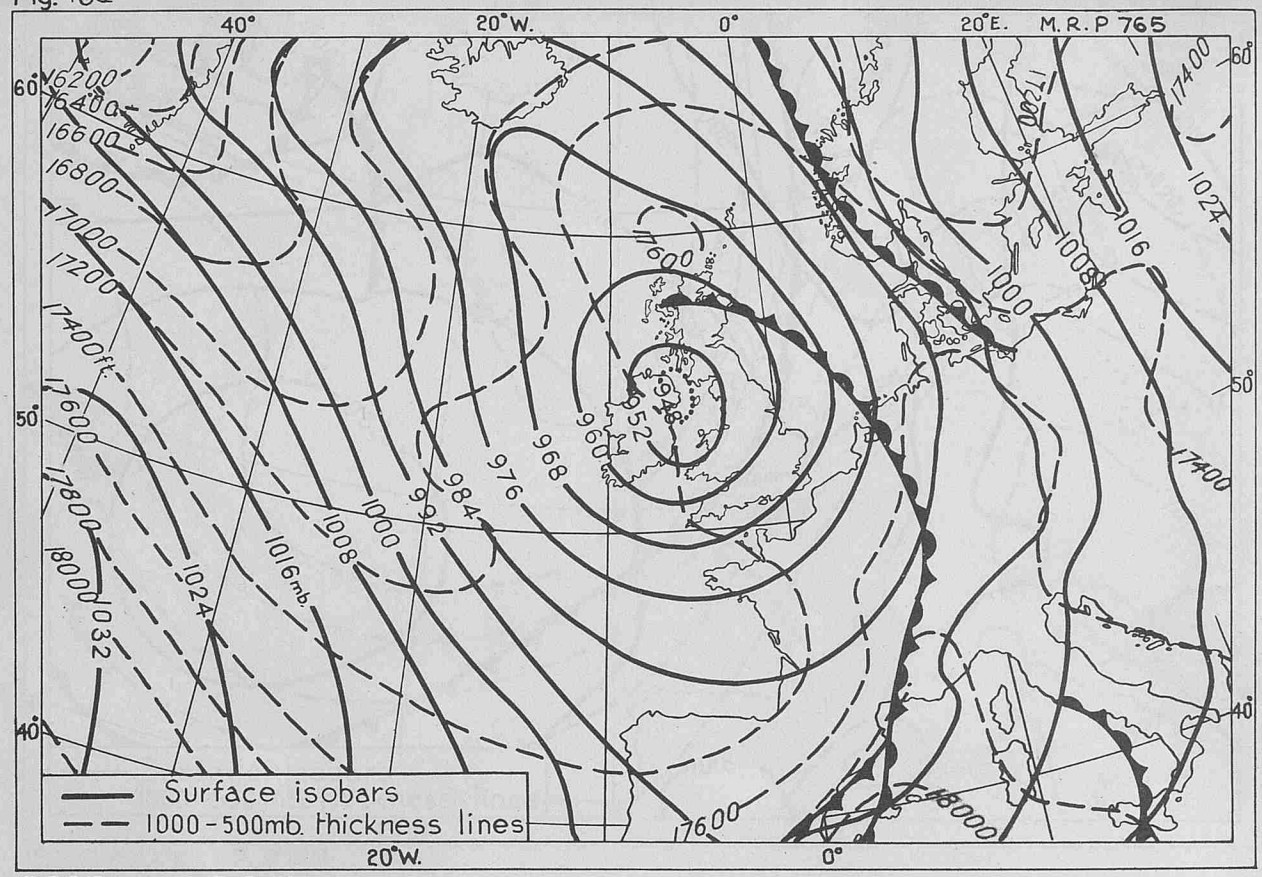


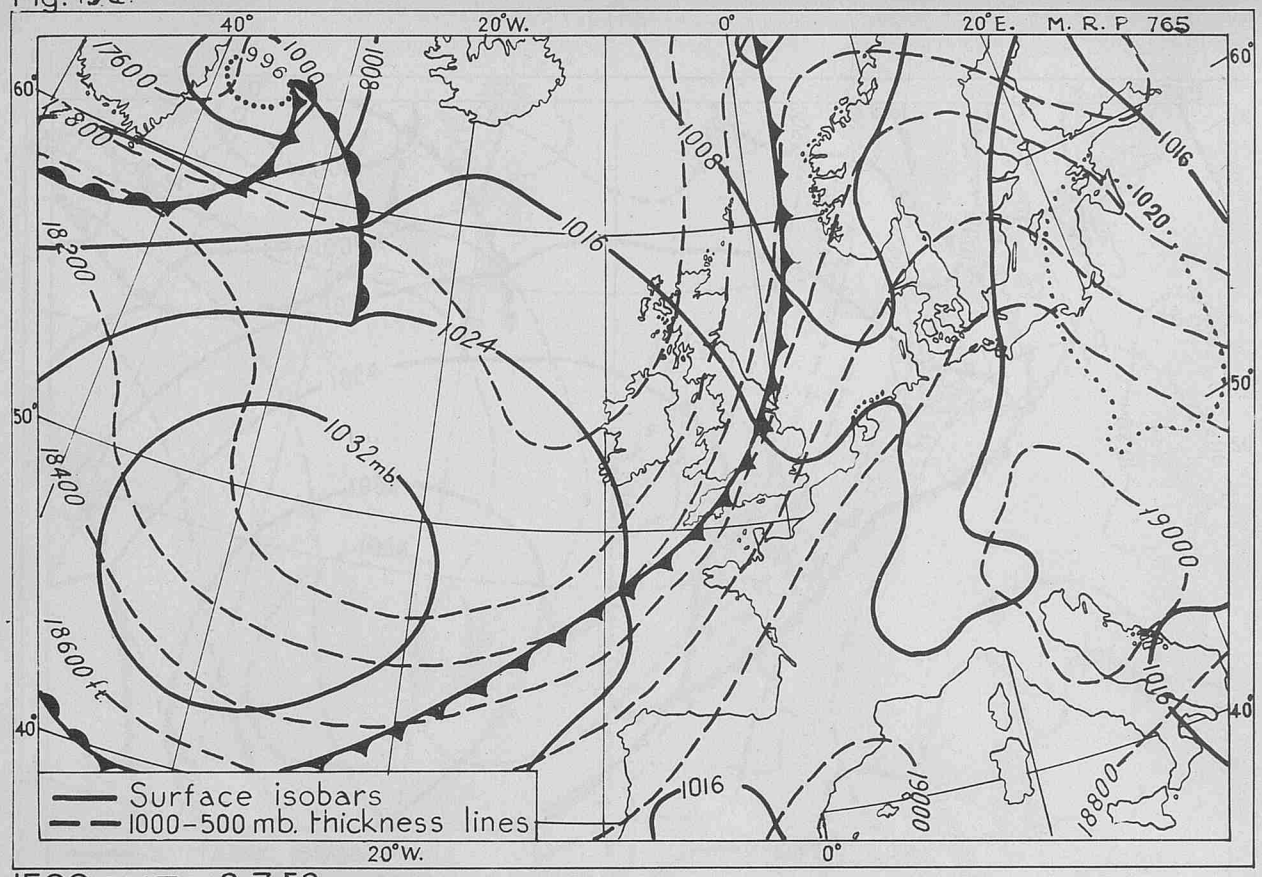
FIG. 17b. DEVELOPMENT CHART. (UNITS 10^{-3} HRS^{-2}) 1500 GMT 4.2.51.

Fig. 18a



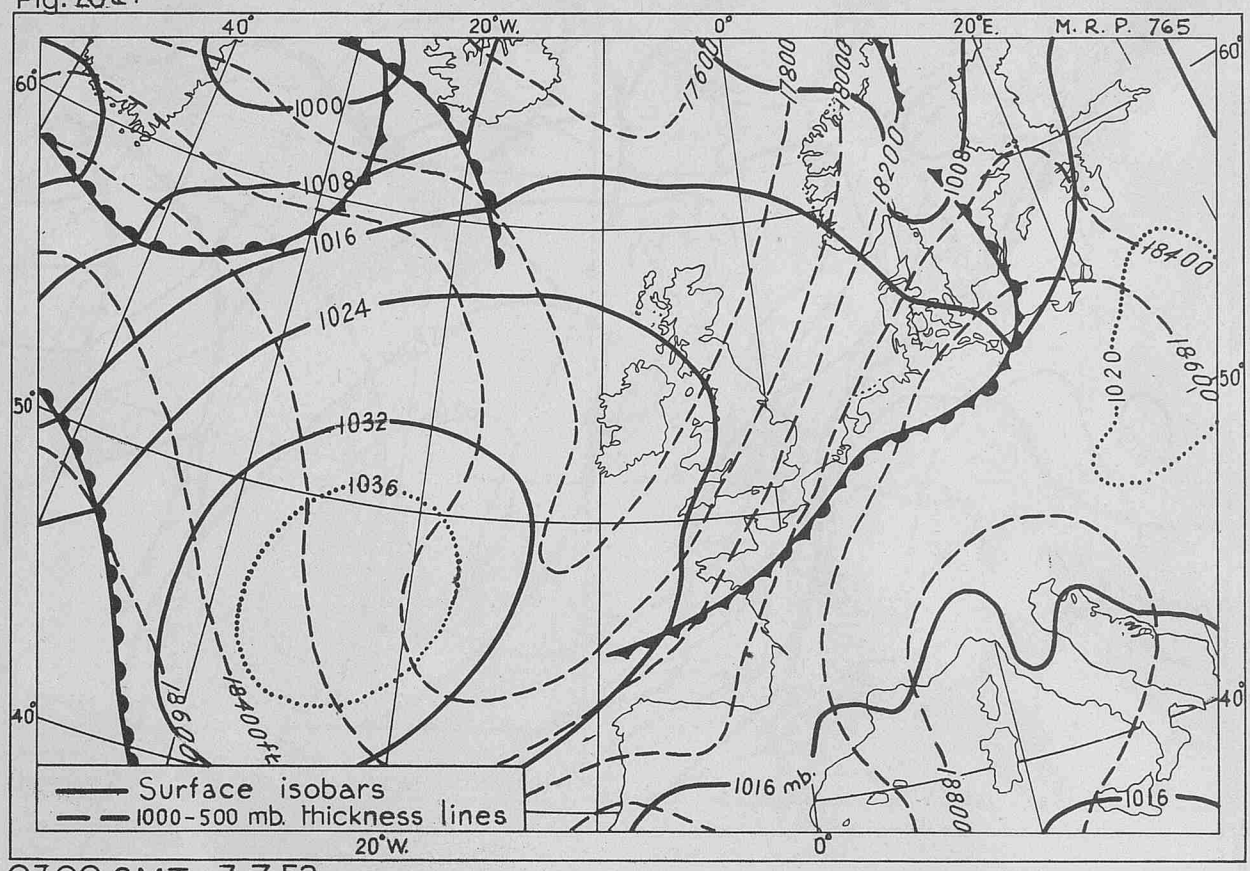
0300 G.M.T. 5.2.51.

Fig. 19a.



1500 G.M.T. 2.7.52.

Fig. 20a.



0300 G.M.T. 3.7.52.

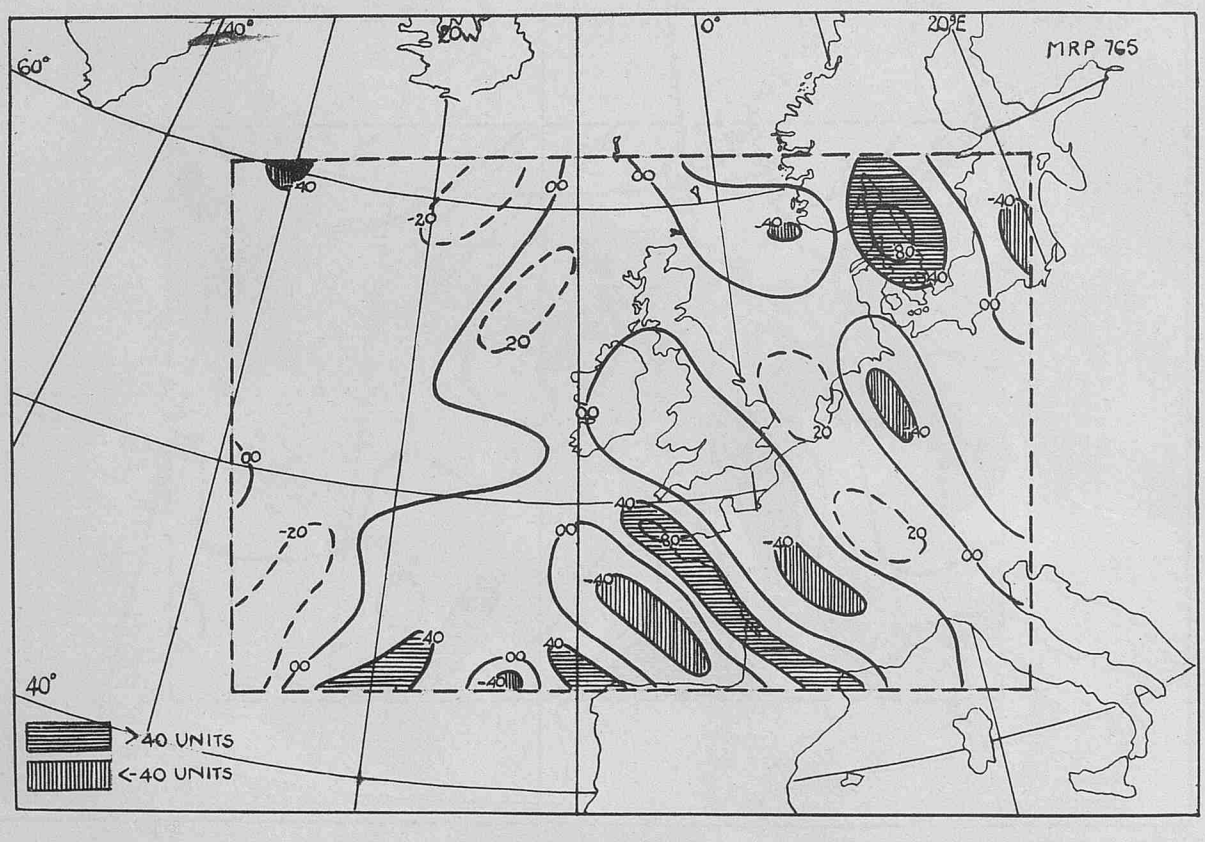
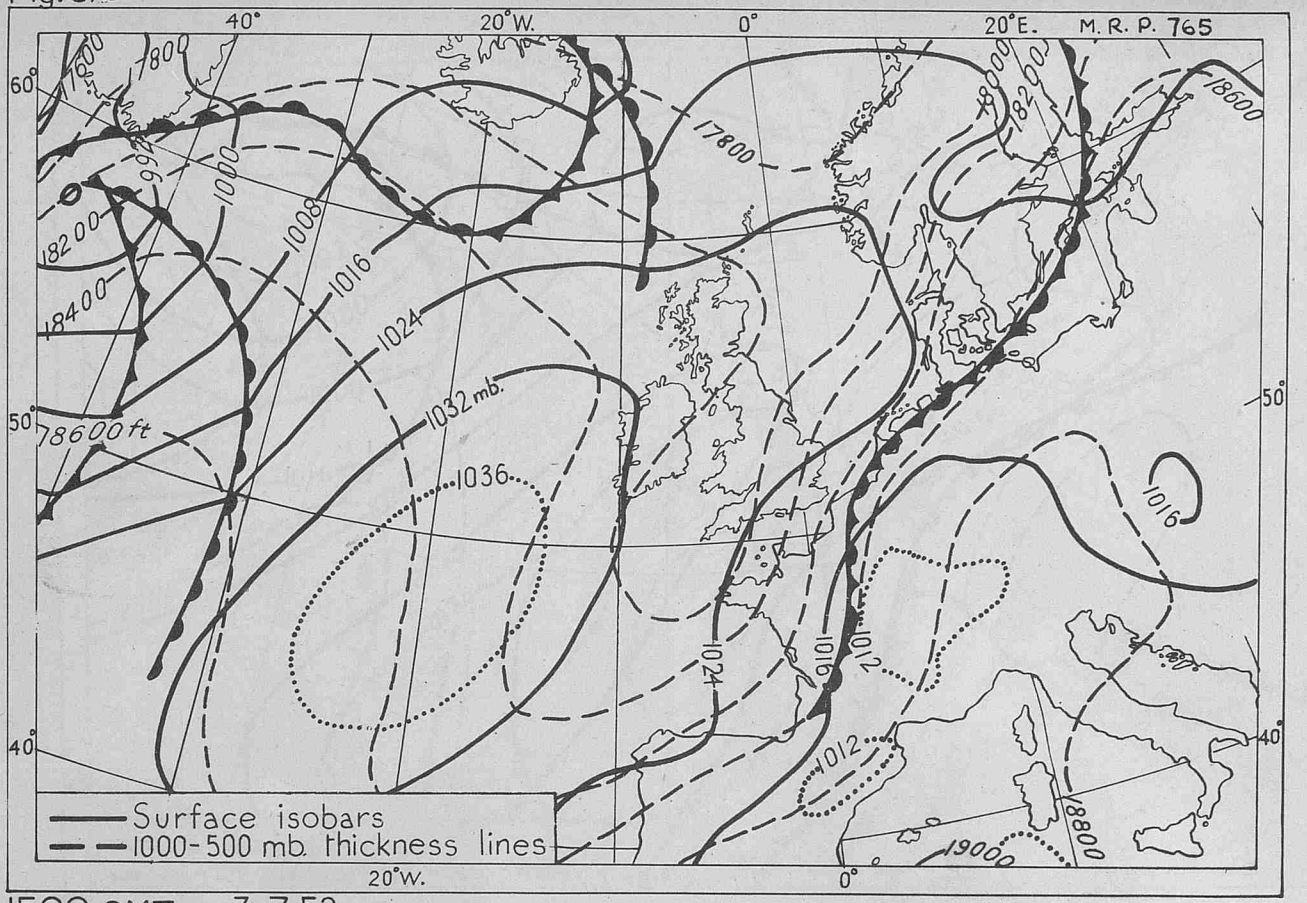


FIG 20b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}). 0300 GMT. 3 · 7 · 52

Fig. 21a.



1500 G.M.T. 3.7.52.

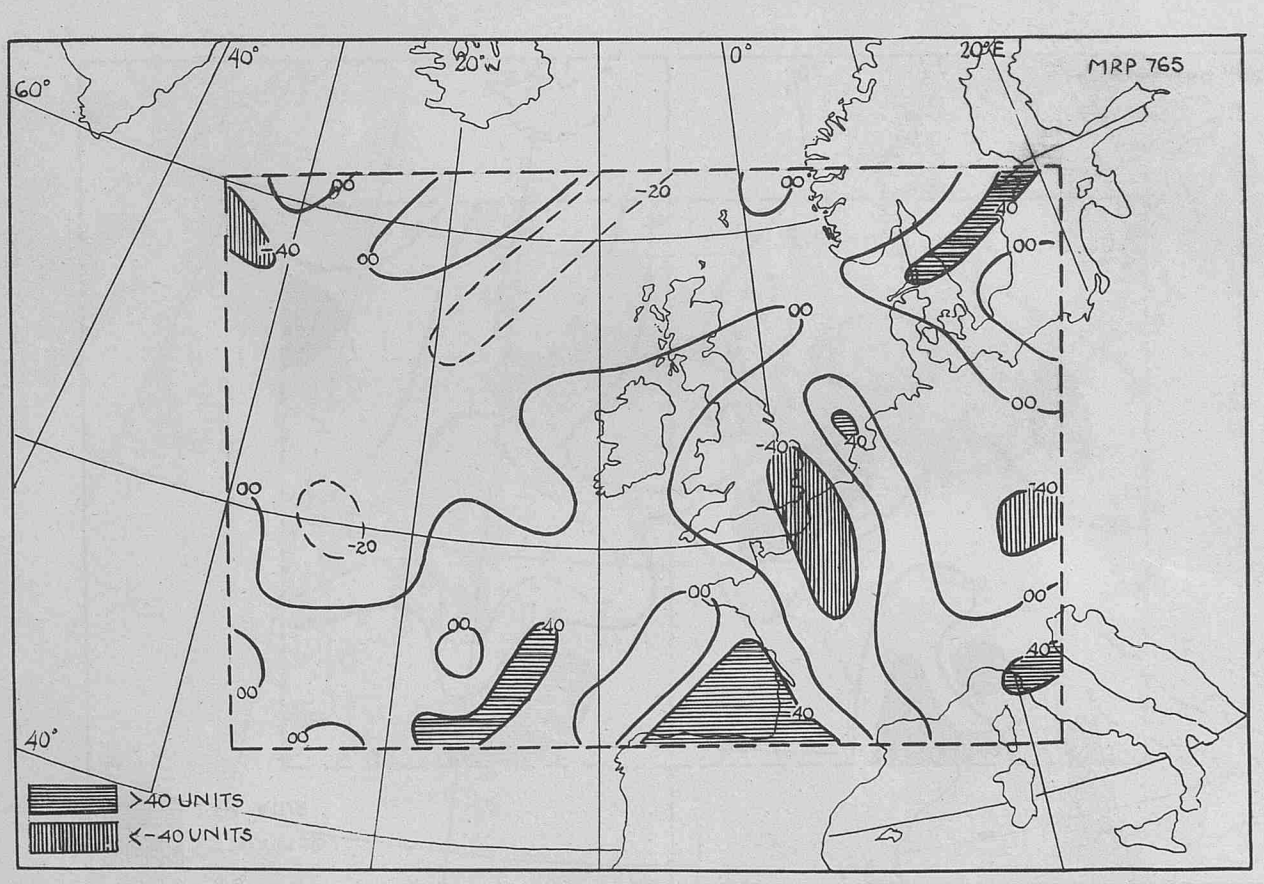
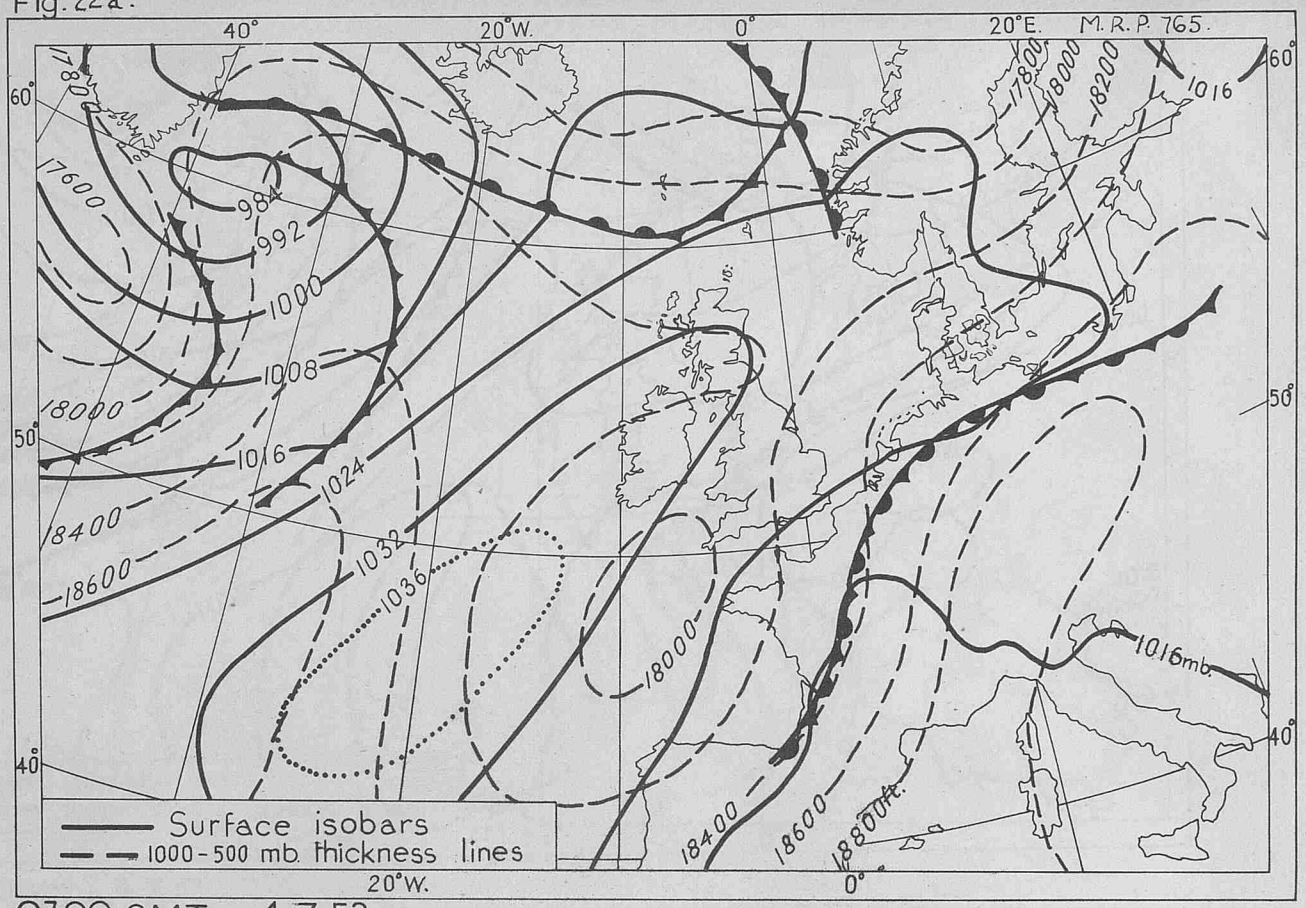


FIG 21b DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 1500 G MT. 3·7·52.

Fig. 22a.



0300 G.M.T. 4.7.52.

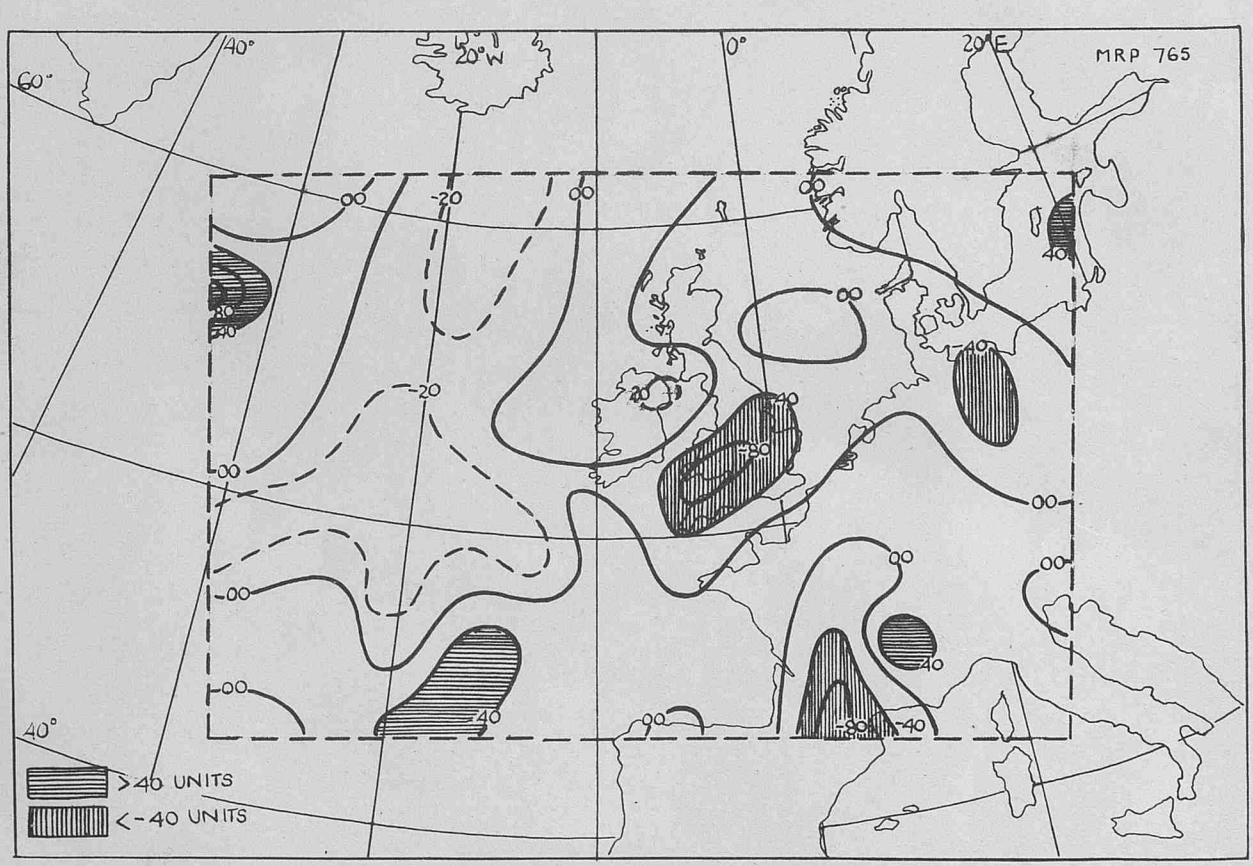
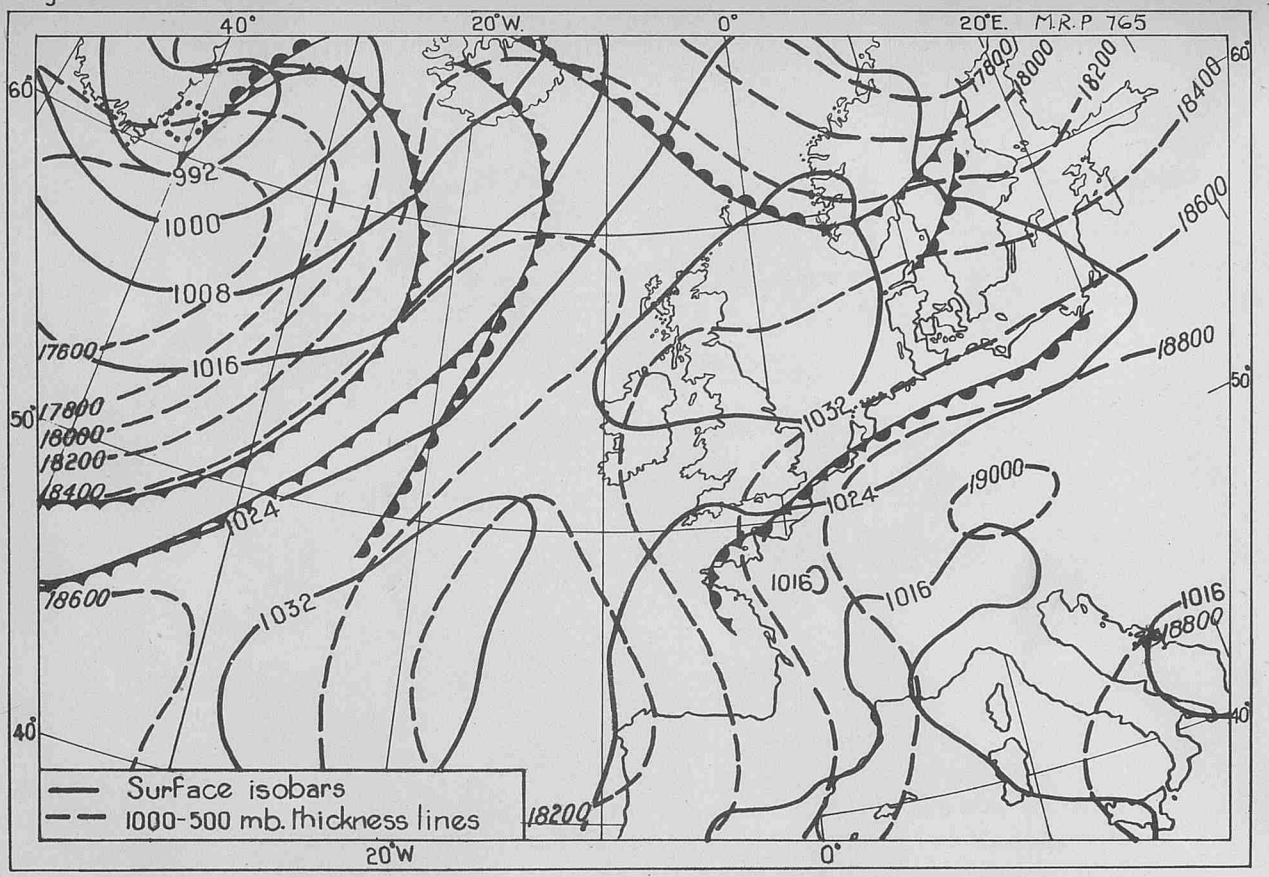


FIG. 22b. DEVELOPMENT CHART (UNITS 10^{-3} HRS^{-2}) 0300 GMT. 4·7·52

Fig. 23 a.



1500 G.M.T. 4.7.52

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M.R.P. 770
S.C.II/130

17th November, 1952.

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Wind at high levels over Singapore (1950-52)

by
R. F. M. Hay (Meteorological Office)

1. Summary.

During the past two years, radar wind finding ascents have been made twice weekly at Singapore using a G.L.III set Type III Mark 2. Data for 149 ascents are available, of which 101 were followed to a height exceeding 55,000 ft. and 18 to over 70,000 ft. Easterlies prevail throughout the year above 25,000 ft., gradually increasing with height almost to the tropopause, a change to south-westerly winds at greater heights occurring between 55,000 and 60,000 ft. in November to February and May to July and between 60,000 and 65,000 ft. in the remaining months. In the period of most active monsoon at Singapore (June to early September, December to mid-March) easterlies at levels between 40,000 and 55,000 ft. become notably stronger than in other months. Winds exceeding 90 kt. have been observed and more than 10% of all ascents gave easterlies exceeding 60 kt. somewhere between these levels. Marked vertical shear was noted and among thirty occasions for which shears were computed two showed positive shear (i.e. wind increasing with height) and nine showed negative shear (i.e. wind decreasing with height) which exceeded 10 kt./1,000 ft. These strong easterlies usually occur in a layer between 48,000 and 54,000 ft. having a superficial resemblance to a "jet stream"; however, no evidence is yet to hand for determining the horizontal extent of these strong easterlies in the vicinity of Singapore.

2. Brief History.

Ascents were begun on 3rd July, 1950 using a radar set (G.L.III Type III Mark 2 manually operated) located on Blakang Mati Island opposite Keppel Harbour, Singapore. The site of the radar set was altered on 24th April, 1951 when it was moved a distance of one mile from Mount Imbia to Bukit Maris. Several changes have had to be made in the location of the balloon release site which are not considered to have affected the results. Up to December, 1950 the G.L.III set was coupled to a predictor on which the target could be followed visually in an incorporated theodolite until it entered cloud. A comparison between the readings of azimuth and elevation obtained visually and by radar showed that the standard of accuracy maintained by the Asiatic service personnel who operated the radar set was generally satisfactory. The other duties associated with radar wind finding, preparation and launching of targets and computations of readings have been carried out by staff of the Meteorological Office with some assistance from Malayan Meteorological Service staff. The local time of ascent was 0800-0930 (0030-0200 G.M.T.). Singapore is just over one degree north of the Equator. Initially one ascent was made weekly with some interruptions. Since 26th January, 1951 there have been two ascents in most weeks.

/3.

G.352157/IW/11/52/40

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3. Mean Winds.

The number of observations available at the higher levels is shown in the table below:-

Height (ft.)	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	1st Sept., 1950 -31st Aug., 1952
75,000	0	0	0	0	0	1	0	0	0	0	0	0	1
70,000	2	0	6	1	4	1	0	0	0	0	2	2	18
65,000	4	6	10	5	6	2	1	1	0	2	4	8	49
60,000	6	6	13	8	9	4	3	4	3	3	5	8	72
55,000	7	7	14	12	11	11	4	9	6	6	6	8	101
50,000	6	10	15	12	11	14	11	12	10	9	9	8	127
45,000	7	12	14	13	11	15	13	17	10	10	9	8	139
40,000	7	12	14	13	11	16	13	18	11	11	9	9	144
35,000	7	13	14	13	11	16	13	20	11	11	9	9	147
30,000	7	13	14	13	11	16	13	20	12	11	9	10	149

Although in most months observations are distributed evenly in time, the number available is not sufficient to justify preparation of wind roses. Vector resultants and average scalar wind speeds have been computed for each 5,000 ft. interval from 10,000 to 65,000 ft. using the period 1st September, 1950 to 31st August, 1952. These are shown in Table I and Fig. 1. (Without high level temperature data it was not considered justifiable to convert height data to pressure levels by adopting a standard atmosphere for Singapore). The typical wind structure disclosed is a rise of the easterly component almost up to the tropopause with a rapid change to light westerlies in the stratosphere. Examples are shown in Fig. 2. The mean wind speeds in Table I for levels above 35,000 ft. much exceed the values found for Batavia and quoted by Shaw (1).

In the upper troposphere at least above 25,000 ft. easterlies prevail throughout the year. The direction in most cases lies near ENE. becoming E. by S. in February and March at 40-45,000 ft. and around NE. in June to August at 40-50,000 ft. Mean monthly speeds show a gradual increase with height from 10 to 15 kt. at 25,000 ft. to 40 to 60 kt. at some levels above 40,000 ft. in certain months. Below 50,000 ft. the winds show a high degree of constancy. The strongest winds are found at 50,000 ft. in most months of the year.

4. Seasonal features.

The most interesting feature of the high level wind regime over Singapore appears to be the existence of strong easterlies which have been observed at levels between 40,000 and 55,000 ft. Wind speeds exceeding 90 kt. have occurred on several occasions and a substantial number of cases of winds exceeding 60 kt. have occurred amounting to $>$ 10% of cases through the whole period. With very few exceptions days on which the highest winds are observed have occurred during the period of fullest development of the monsoon at Singapore, i.e. from June to early September and December to mid-March. This is confirmed by figures in the table of vector resultants, and by an examination of the variation through the year of the values of extreme wind speeds reached on each ascent, irrespective of level, for the ascents which exceeded 55,000 ft. Both methods show that the strongest winds occur at high levels in January and February, also in July and August, the latter months appearing to yield the higher speeds. The main features of the wind distribution are similar in both years.

In the absence of radio-sonde ascents there are at present no data for upper air temperatures over Singapore for levels above 30,000 ft. Data available for other tropical localities together with data from radio-sonde ascents recently

/started

started in Ceylon suggest that the tropopause over Singapore is usually at an elevation of 55,000 - 60,000 ft. The table of vector resultants shows an interesting change of wind with height which in the pre- and early monsoon months (November-February and May-July) occurs slightly below 60,000 ft. and in the late monsoon and transition months (March-April, August-October) probably occurs at or slightly above 60,000 ft.

The very few ascents which reached 65,000 to 75,000 ft. indicate south-westerly winds in this layer in most months.

Mean annual temperatures at standard pressure levels over Singapore given by John (2) have been used to compute mean heights of pressure surfaces over Singapore. The results are given below for interest to assist comparison with results for other localities which relate to pressure heights. As no temperature data were available for Singapore above 300 mb. it was decided to use data from a few radiosonde ascents from Colombo which have recently become available, for making estimates of mean temperatures for Singapore at levels above 300 mb.

Pressure	Mean height of pressure surface (ft.)	Pressure	Mean height of pressure surface (ft.)
100 mb.	54,600	300 mb.	31,700
150	46,600	350	28,200
200	40,700	400	24,800
250	35,700		

5. Some features of days with strong easterlies.

The heights at which strong easterlies occur have been investigated. In the monsoon months the strongest winds occur in the layers 48,000 to 54,000 ft., during the transition months between 46,000 to 52,000 ft. Winds in these layers average 15-20 kt. stronger in the monsoon months than at other times. Table III gives the distribution in detail. On many days of strong easterlies the balloon evidently passed out of range at lower heights than those to which it was followed on days of lighter wind.

The highest wind speed recorded on an ascent was 95 kt., and on three ascents speeds over 90 kt. have occurred. On nine occasions over 80 kt. was recorded, six of these on ascents which did not reach 55,000 ft. This makes it probable that even stronger winds would have been observed on some of the ascents which terminated some thousands of feet below the 55,000 ft. level, or else that some readings were inaccurate. Accordingly a check of the 'irregularity' of rate of ascent of balloons was made. 'Irregularity' was defined as the percentage of the total number of height intervals on an ascent which exceeded more and less than (a) 10%, (b) 25% of the value of the mean rate of ascent per minute. These quantities were readily obtained since the readings used here were all made at one minute intervals. A series of five days was selected in each year for occasions of both strong and light easterlies, making 20 days in all. In addition seven different days on which the strongest winds of all were recorded were used as a check. The results showed:

(a) In the first year there was no difference in "irregularity" in rate of ascent between days of strong and light winds.

(b) In the second year days of light wind show more "irregularity" than days of strong wind.

(c) For seven selected cases of wind extremes between 70 and 90 kt. "irregularity" was less than the mean irregularity for the 20 cases ("all winds") over the two years.

(d) The mean rates of ascent were:-

10 cases of "light" winds	1,070 ft. p. sec.
10 " " "strong" "	1,230 " "
7 " " "strongest" winds	1,300 " "

/The

The first three results were unexpected but all indicate that the readings can be relied upon. The last result is to be expected.

A comparison was made between the frequency distributions of maximum wind speeds noted on each ascent in the two twelve month periods 1st September, 1950 - 31st August, 1951 and 1st September, 1951 - 31st August, 1952. Both for ascents ending above 55,000 ft. and ending below 55,000 ft. the agreement between the two years was close. Slightly higher frequencies of winds in the 10kt. wind ranges over and above 61-70 kt. in the second year can be partly attributed to a greater proportion of ascents having been made during the monsoon months in the second year than in the first year, and partly to a true higher frequency of maximum winds over 60 kt. which occurred in the monsoon months of the second year.

6. Shears.

From the variation of the seasonal mean vector winds with height it is evident that vertical shear of considerable magnitude must often occur just below the probable level of the tropopause over Singapore. A number of days with strong easterlies at high levels were examined and measurements of the largest shears on each occasion are shown in Table IV which separates days on which ascents were made to a height exceeding 55,000 ft. from those on which ascents terminated at a lower level. Shears were computed in knots per 1,000 ft. by direct computation from the vector wind shear through the layer obtained from the wind speeds at the base and top of the layers selected on the reasonable assumption that the wind components varied smoothly through the layers. The thickness of the layers was arbitrarily restricted to between 2,000 and 5,000 ft. though in practice most of the layers used were $>$ 3,000 ft. thick. This procedure yields mainly higher values of shear than would be given by smoothing and resolving along a particular direction, but the comparison between the two procedures which is given later shows that the differences are not as a rule significant. This is because the layers were carefully selected from ascents which were considered reliable after examination. A large number of the cases quoted were examined graphically (wind velocity plotted against height) and any ascents with readings suspected as being inaccurate through errors in reading, or timing of readings, have been excluded.

The criterion of a minimum shear of 5 kt./1,000 ft. has been adopted for all occasions of "strong wind shear", since it agrees well with the statement made by several workers that a vertical shear of 80 kt./12,000 ft., (i.e. 6.7 kt./1,000 ft.) is often met in "jet streams" in temperate latitudes.

Table IV shows that 58 ascents examined in this part of the work "caught" 8 cases of positive shear and 12 cases of negative shear up to the greatest heights reached, while a further 10 cases of positive and 3 cases of negative shears were "caught" in ascents terminated below 55,000 ft. Almost all the 22 ascents included in this Table show the value of negative shear on the upper side of the strong wind stream to be much larger than the associated positive shear on the lower side. In general the values of negative shear recorded are considerably larger than the values of positive shear, 10 out of the total of 15 negative shear values exceed 10 kt./1,000 ft. while 2 exceed twice this value. Out of a total of 18 positive shear values, only 2 exceed a value of 10 kt./1,000 ft.

Although the layers used for measuring shears for Table IV were selected as having small changes of wind direction through their thickness, it was considered advisable to make a comparison for a few cases between shears measured directly from wind speeds (ignoring change of direction), and shears measured from components of wind resolved along a direction closely parallel to the mean wind direction for each separate layer. As wind components were computed along the direction of flow of air at the level of highest wind speed for another purpose the comparison was made for these days as given below:-

/Shears

1951	Shears computed from wind speeds +	Shears computed from wind speeds −	Shears computed from wind components +	Shears computed from wind components −
6 Feb.	11.0	...	8.2	...
9 Feb.	5.8	21.7	4.8	16.0
11 May	7.6	11.3	8.7	16.0
29 June	(8.1	14.2	(7.4	14.4
	(8.9	13.5	(7.6	7.6

The agreement is satisfactory. Larger values would be expected for the first method, which overestimates the shear where wind direction changes with height and because estimates made graphically from the wind components were made conservatively and for rather thicker layers.

The extreme values of shears so far computed (+15.6 kt./1,000 ft. on 3rd August, 1950 and -21.7 kt./1,000 ft. on 9th February, 1951) are similar in magnitude to values quoted by Bannon (3) in M.R.P. 631 (S.C.III/92) dated 1st February, 1951 and in earlier work. In temperate latitudes shears of this magnitude are frequently associated with jet streams. However, in the absence of any evidence of marked horizontal thermal gradients in the easterlies over Singapore, along with the negligible value of the Coriolis force, it is evident that the production and maintenance of strong easterlies cannot be explained as being due to the operation of some of the factors which have been assumed to be responsible for, or are found in association with, temperate jet streams. The relationship of the easterlies with other factors is being investigated.

As stated elsewhere no information is yet available for Singapore on lapse rates of temperature above 30,000 ft. The shears described by Bannon on 2nd and 7th November, 1950 were associated with exceptional turbulence at high altitude, some of the cases recorded being a short distance below the tropopause. Most of the cases of negative shear occurred in the layers between 40,000 and 55,000 ft. and more frequently at the higher level. When a start was made with the examination of this series of wind observations it soon became evident from analogy with the findings in the European area, that moderate or severe turbulence in clear air was to be expected at these altitudes in equatorial regions. However, confirmation of this was lacking until publication of a note by Batty (4) recently, in which he states a jet aircraft reported "moderate turbulence from Ceylon to Sabang (N.W. Sumatra) at about 45,000 ft. when flying in the same plane as 2 oktas banded cirrocumulus, for periods of 3-4 min. at 10- min. intervals, the effect on the aircraft suggesting turbulence rolls continuous with cirrocumulus rolls seen off track with their axes at right angles to the track". On this date easterlies between 60 and 90 kt. were found between 40,000 and 45,000 ft. over Singapore.

7. Some additional features of strong wind streams.

The mid points of layers used in computing the shear values quoted lay for positive shears most frequently between 36,000 and 48,000 ft. with the maximum frequency between 40,000 and 48,000 ft. For negative shears the corresponding values were 44,000 and 56,000 ft. with the maximum frequency between 46,000 and 54,000 ft. Cases of two separate layers of strong wind streams below 55,000 ft. were observed on 31st August, 1950 and 15th June, 1951; also on 29th June, 1951, there were two strong wind streams, the lower with maximum wind speed of 060°/70 kt. at 47,000-49,000 ft., the upper with a maximum wind speed of 273°/50 kt. at 59,000 ft., presumably within the stratosphere.

A rapid check has been made on the relative frequency of strong winds at high levels on all days available during the year considered here by taking out the frequencies of various wind speeds at 50,000 ft. for the 127 ascents which reached this height. Results also expressed as percentage frequencies are given in Table V. The vertical thickness of the strong easterlies was examined and results are given in Table VI. The most significant facts are that 8% of the 149 ascents examined showed winds in excess of 60 kt. through a thickness

/between

between 4,000 and 6,000 ft., while 1% of ascents gave both winds exceeding 60 kt. through a thickness between 8,000 and 10,000 ft., and winds exceeding 80 kt. through a thickness between 2,000 and 4,000 ft. The method of measurement gave minimum values for thicknesses and these figures are conservative estimates in most cases; in particular the table shows that in a large number of ascents the upper limit of the strong wind layer was not reached.

Strong easterlies being narrower (vertically) and associated with much lower velocities than jet streams in temperate latitudes are evidently less important as a navigational hazard to jet aircraft than their temperate latitude counterparts. However, although an aircraft flying eastwards near the equator can easily get out of the strongest winds at these high levels by climbing or descending a few thousand feet (a smaller vertical distance as a rule than in temperate latitudes) a reference to Table IV will show that the vertical separation between the levels of maximum shear lies anywhere between 2,000 and 8,000 ft. and most usually is around 5,000 ft. On these days there will evidently be a zone around this height range which should preferably be avoided by aircraft flying in an easterly direction. Information about the horizontal length or breadth of these strong easterlies is lacking.

8. Acknowledgment.

Assistance given by staff of the Meteorological Office serving on Singapore Island (and by some of the Malayan Meteorological Service staff, by courtesy of their Director) in carrying out this radar wind finding programme is gratefully acknowledged.

References.

1. Shaw, N. Manual of Meteorology, Vol. II, p.272.
2. John, I.G. Memoirs of the Malayan Meteorological Service, No. 3.
3. Bannon, J.K. Severe clear air turbulence experienced at high altitude on 2nd and 7th November, 1950. M.R.P.631, 1st February, 1951.
4. Batty, R.P. Meteorological observations on a flight by a Canberra jet aircraft from England to Australia. Met. Mag., Vol. 81, No. 961, July, 1952, p.219.

Table I. VECTOR RESULTANT WINDS AT SINGAPORE.

HEIGHTS a.m.s.l. (feet)	JAN.	FEB.	MAR.	APR.	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	DEC.-FEB.	MAR.-MAY	JUNE-AUG.	SEPT.-NOV.	HEIGHTS (feet)
75,000	-	-	-	-	-	220/18 (1)	-	-	-	-	-	-	-	-	220/18 (1)	-	75,000
70,000	169/07 (2)	-	276/13 (6)	086/46 (1)	251/23 (4)	273/13 (1)	-	-	-	-	045/14 (2)	088/13 (2)	115/08 (4)	262/11 (11)	273/13 (1)	045/14 (2)	70,000
65,000	253/14 (4)	009/04 (6)	218/03 (10)	204/13 (5)	290/19 (6)	254/20 (2)	222/13 (1)	232/24 (1)	-	286/20 (2)	265/19 (4)	260/15 (8)	266/09 (18)	254/08 (21)	242/19 (4)	272/19 (6)	65,000
60,000	259/16 (6)	315/07 (6)	069/05 (13)	037/10 (8)	286/04 (9)	268/21 (4)	029/11 (3)	049/11 (4)	355/14 (3)	019/07 (3)	263/12 (5)	248/09 (8)	265/09 (20)	038/04 (30)	328/04 (11)	315/07 (11)	60,000
55,000	075/16 (7)	094/15 (7)	076/04 (14)	055/16 (12)	079/11 (11)	070/19 (11)	079/27 (4)	091/24 (9)	070/24 (6)	075/06 (6)	078/16 (6)	061/12 (8)	078/14 (22)	066/10 (37)	080/22 (24)	074/16 (18)	55,000
50,000	100/35 (6)	099/42 (10)	078/10 (15)	064/14 (12)	057/25 (11)	070/37 (14)	069/60 (11)	077/57 (12)	077/29 (10)	058/28 (9)	084/32 (9)	082/26 (8)	094/35 (24)	064/15 (38)	072/50 (37)	074/30 (28)	50,000
45,000	106/33 (7)	106/41 (12)	119/18 (14)	081/14 (13)	063/23 (11)	065/43 (15)	062/47 (13)	068/49 (17)	071/34 (10)	065/27 (10)	078/23 (9)	086/20 (8)	103/32 (27)	088/16 (38)	065/47 (45)	070/28 (29)	45,000
40,000	104/26 (7)	110/31 (12)	114/21 (14)	076/15 (13)	053/13 (11)	068/37 (16)	057/39 (13)	066/41 (18)	064/26 (11)	071/20 (11)	081/15 (9)	085/19 (9)	103/26 (28)	088/15 (38)	064/39 (47)	070/21 (31)	40,000
35,000	098/17 (7)	111/23 (13)	109/18 (14)	074/16 (13)	069/09 (11)	068/28 (16)	066/29 (13)	069/27 (20)	071/21 (11)	070/15 (11)	083/14 (9)	075/10 (9)	102/16 (29)	089/14 (38)	068/28 (49)	073/16 (31)	35,000
30,000	093/16 (7)	103/16 (13)	098/12 (14)	067/15 (13)	060/05 (11)	078/23 (16)	082/24 (13)	083/20 (20)	078/20 (12)	092/12 (11)	097/11 (9)	041/08 (10)	089/12 (30)	080/11 (38)	081/22 (49)	086/14 (32)	30,000
25,000	084/14 (7)	078/11 (13)	073/09 (14)	082/12 (13)	080/03 (10)	091/18 (15)	092/16 (13)	091/16 (20)	077/17 (12)	099/10 (11)	090/07 (9)	007/03 (10)	074/09 (30)	079/08 (37)	091/17 (48)	085/11 (32)	25,000
20,000	068/09 (7)	068/09 (13)	066/05 (15)	079/10 (13)	345/01 (10)	093/06 (16)	120/04 (13)	108/05 (20)	079/09 (12)	095/07 (11)	086/07 (9)	009/01 (10)	065/06 (30)	072/06 (38)	104/05 (49)	086/08 (32)	20,000
15,000	070/01 (7)	046/04 (13)	360/01 (15)	098/04 (13)	308/04 (11)	261/05 (16)	228/04 (13)	225/04 (20)	047/03 (12)	128/04 (10)	049/05 (9)	031/02 (10)	045/01 (30)	045/01 (39)	239/04 (48)	071/03 (31)	15,000
10,000	271/09 (6)	328/03 (12)	245/07 (13)	081/05 (11)	293/03 (16)	253/12 (13)	259/09 (22)	257/09 (12)	238/05 (10)	162/04 (10)	070/01 (8)	028/01 (10)	298/03 (28)	260/03 (35)	256/10 (51)	203/02 (30)	10,000

Explanation:- 271/09 ... wind direction 271° true, speed 9 knots
(6) .... figure below in brackets denotes the number of observations

- ... dash denotes no observations

G.352157(a)/IW/11/52/40

TABLE II. SEASONAL VALUES OF AVERAGE SCALAR WIND SPEED (KNOTS)

AT SINGAPORE.

Heights (feet)	Dec.- Feb.	Mar.- May	June- Aug.	Sept.- Nov.
75,000	-	-	18 (1)	-
70,000	19 (4)	24 (11)	13 (1)	16 (2)
65,000	15 (18)	17 (21)	20 (4)	20 (6)
60,000	18 (20)	15 (28)	21 (9)	15 (11)
55,000	25 (22)	20 (35)	24 (17)	21 (18)
50,000	37 (24)	23 (36)	53 (37)	31 (28)
45,000	35 (27)	22 (36)	48 (45)	30 (29)
40,000	27 (28)	20 (36)	41 (47)	21 (32)
35,000	20 (29)	17 (36)	30 (49)	18 (31)
30,000	17 (30)	15 (36)	24 (49)	16 (32)
25,000	13 (30)	11 (35)	19 (48)	13 (32)
20,000	14 (30)	9 (36)	10 (49)	11 (32)
15,000	11 (30)	10 (37)	12 (49)	11 (31)
10,000	12 (28)	11 (33)	14 (49)	12 (30)

Figures in brackets give number of observations.

TABLE III. DISTRIBUTION WITH HEIGHT OF HIGHEST WIND SPEED RECORDED ON EACH ASCENT AT SINGAPORE DURING PERIOD 1st SEPTEMBER, 1950 TO 31st AUGUST, 1952.

Heights (100 s ft.)	Ascents reaching and exceeding 55,000 ft.		Ascents not reaching 55,000 ft.	
	Monsoon months	Transition months	Monsoon months	Transition months
> 571	Nil	Nil	Nil	Nil
551–570	57 (1)	35 30 28 (3)		
531–550	60 60 48 39 62 36 47 (7)	38 22 40 27 24 (5)	54 (1)	
511–530	49 45 90 44 45 72 60 65 (8)	83 25 30 54 34 30 37 31 42 33 (10)	84 (1)	36 49 (2)
491–510	48 29 60 53 37 72 33 70 66 (9)	30 48 40 46 42 25 49 30 (8)	90 87 63 84 (4)	46 (1)
471–490	70 66 69 78 60 42 54 (7)	42 53 24 24 (4)	66 75 70 85 78 72 (6)	71 66 58 (3)
451–470	29 60 54 (3)	57 52 46 47 (4)	63 83 54 78 (4)	54 27 36 (3)
431–450	40 95 42 40 (4)	30 51 37 (3)	52 43 69 66 67 (5)	28 42 (2)
411–430	30 (1)	26 53 23 42 37 (5)	54 (1)	
391–410	(0)	23 26 (2)	45 41 69 60 63 (5)	47 (1)
371–390	53 44 (2)	30 40 24 (3)	49 32 (2)	
351–370	(0)	49 18 (2)		
< 350	35 29 (2)	38 26 30 21 (4)	18 28 (2)	27 20 (2)
Total occasions	(44)	(53)	(31)	(14)

Each entry denotes the highest wind speed (in knots) on an ascent.

Figures in brackets denote number of occasions in each category.

Monsoon months:- Dec. Jan. Feb. June, July, Aug.
Transition months:- Mar. April, May, Sept. Oct. Nov.

TABLE IV

Shear values computed from Singapore R/W ascents.

Occasions when ascent reached/exceeded 55,000 ft. and strong shear (> 5 Kt/1,000 ft.) was observed.

Date	Heights between which shear was measured (100s feet)	Vertical Shear (Kt/1,000 ft.)		Extreme wind on Ascent Height (100s feet)	Extreme wind on Ascent Dir/Vel (Kt)	Terminal Height of Ascent (100s ft)
1950						
24 Aug.	476-506		-6.3	(415 (476	068/48 070/48	569
31 Aug.	439-467		-13.6	439	046/51	551
29 Sept.	307-355	+5.8		355	079/49	(595)
9 Nov.	449-494	+6.7		527	095/83	665
1951						
8 May	(446-483 (497-531	+9.5	 -6.5	(483 (497	063/42 072/42	616
11 May	(490-531 (544-574	+7.6	 -11.3	531	090/40	691
15 May	440-490		-5.7	440	099/51	683
18 May	525-554		-11.7	497	070/46	554
25 May	458-478		-20.3	458	092/46	724
12 June	525-549		-14.7	507	048/48	571
15 June	(341-383 (388-416 (508-542 (550-584	+6.6 +5.3 	 -9.6 -11.1	388	061/53	584
29 June	(424-471 (490-514 (547-589 (607-635	+8.1 +8.9 	 -14.2 -13.5	490	060/70	635

Occasions when ascents finished below 55,000 ft. and strong shear (> 5 Kt/1,000 ft.) was observed.

Date	Heights between which shear was measured (100s feet)	Vertical Shear (Kt/1,000 ft.)		Extreme wind on Ascent Height (100s feet)	Extreme wind on Ascent Dir/Vel (Kt)	Terminal Height of Ascent (100s ft)
1950						
27 July	380-425	+6.7		425	071/93	434
3 Aug.	(403-428 (428-449	+15.6	 -12.9	428	076/96	(521)
10 Aug.	385-430	+6.4		466	074/94	(466)
7 Sept.	415-463	+5.2		480	070/71	541

Occasions when ascents finished below 55,000 ft. and strong shear (> 5Kt/1,000 ft.) was observed. (contd.)

Date	Heights between which shear was measured (100s feet)	Vertical Shear (Kt/1,000 ft.)	Extreme wind on Ascent		Terminal Height of Ascent (100s ft)
			Height (100s feet)	Dir/Vel (Kt)	
1951					
6 Feb.	438-480	+11.0	496	113/90	522
9 Feb.	(411-449 (487-517	+5.8 -21.7	449	115/75	517
16 Feb.	368-400	+5.3	(421 (449 (470	121/54 131/54 124/54	470
27 Feb.	439-474	+6.0	474	114/70	489
18 June	437-484	+8.3	484	074/78	(522)
25 June	(353-394 (430-473	+5.4 -7.1	403	064/45	473

Shear values computed from wind components resolved along axes parallel to strong easterlies.

Date	Mid-level of shear layer (100 s ft).	Vertical shear (Kt/1,000 ft).	
		+	-
18 July	470	11.2	
	520		11.4
20 July	495	10.2	
	545		11.0
2 Aug.	225	8.0	
	370	6.6	
3 Aug.	360	4.6	
	405		10.0
	420	18.8	
7 Aug.	380	6.0	
	445	4.5	
	515		14.0
	580	8.2	
	620		16.0
13 Aug.	350	6.7	
	415		13.2
17 Aug.	435	13.6	
17 Sept.	460	12.8	

FREQUENCY DISTRIBUTION OF WIND SPEED AND DIRECTION AT THE 50,000 FT. LEVEL OVER SINGAPORE
1st SEPTEMBER, 1950-31st AUGUST, 1952.

TABLE V.

Number of Occasions

Wind Speed (Knots)	0	1-10	11-20	21-30	31-40	41-50	51-60	61-70	71-80	81-90	90	Total
Monsoon Months (Dec. Jan. Feb. June, July, Aug.)	0	2	6	4	13	9	8	11	5	3	0	61
Transition Months (Mar. Apr. May, Sept. Oct. Nov.)	0	7	19	16	11	10	1	2	0	0	0	66
All Months	0	9	25	20	24	19	9	13	5	3	0	127
All Months (as percentage)	0	7	20	16	19	15	7	10	4	2	0	100

Number of Occasions

Wind Direction (°True)	011-030	031-050	051-070	071-090	091-110	111-130	131-150	151-170	171-190	191-210	211-230
Monsoon Months	1	6	12	20	15	3	1	0	0	1	0
Transition Months	5	7	17	17	8	0	2	1	0	0	2
All Months	6	13	29	37	23	3	3	1	0	1	2
All Months (as percentage)	5	10	23	28	18	2	2	1	0	1	2

	231-250	251-270	271-290	291-310	311-330	331-350	351-010	Calm	Total
Monsoon Months	2	0	0	0	0	0	0	0	61
Transition Months	1	1	1	1	1	1	1	0	66
All Months	3	1	1	1	1	1	1	0	127
All Months (as percentage)	2	1	1	1	1	1	1	0	100

FREQUENCY OF OCCURRENCE OF STRONG WINDS THROUGH WHOLE LAYERS OF VARIOUS THICKNESS. TABLE VI.
SINGAPORE, 1st SEPTEMBER, 1950 to 31st AUGUST, 1952.

Layer thickness (feet)	Ascents reaching and exceeding 55,000 ft.			Ascents ending below 55,000 ft.			All ascents			Frequency of occurrence as percentage of total number of ascents.		
	Wind speed			Wind speed			Wind speed			Wind speed		
	≥40	≥60	≥80	≥40	≥60	≥80	≥40	≥60	≥80	≥40	≥60	≥80
< 2,000*	38(1)	17	3	29(18)	21(10)	6(0)	67(19)	38(10)	9(0)	45	25	6
> 2,000	33(1)	10	1	28(18)	15(9)	0(0)	61(19)	25(9)	1	41	17	1
> 4,000	26	5	0	25(17)	7(6)	0	51(17)	12(6)	0	34	8	0
> 6,000	18	3	0	20(16)	3(2)	0	38(16)	6(2)	0	25	4	0
> 8,000	10	1	0	14(11)	1(1)	0	24(11)	2(1)	0	16	1	0
>10,000	7	0	0	10(7)	0(0)	0	17(7)	0(0)	0	11	0	0
>12,000	6	0	0	6(4)	0	0	12(4)	0	0	8	0	0
>14,000	5	0	0	3(3)	0	0	8(3)	0	0	5	0	0
>16,000	3	0	0	1(1)	0	0	4(1)	0	0	3	0	0
>18,000	2	0	0	1(1)	0	0	3(1)	0	0	2	0	0
>20,000	1	0	0	0(0)	0	0	1(0)	0	0	1	0	0
>22,000	0	0	0	0	0	0	0(0)	0	0	0	0	0
	*For 4 cases winds in layer ≥ 60kt. and 2 cases winds ≥ 80kt. this speed was reached at only one observed level.			*For 2 cases winds ≥ 60 kt. and 4 cases winds ≥ 80 kt. this speed was reached at only one observed level.			Figures in brackets give the number of occasions when strong winds persisted up to the level of termination of the ascent and the true thickness of the layer of strong winds could not be determined.					

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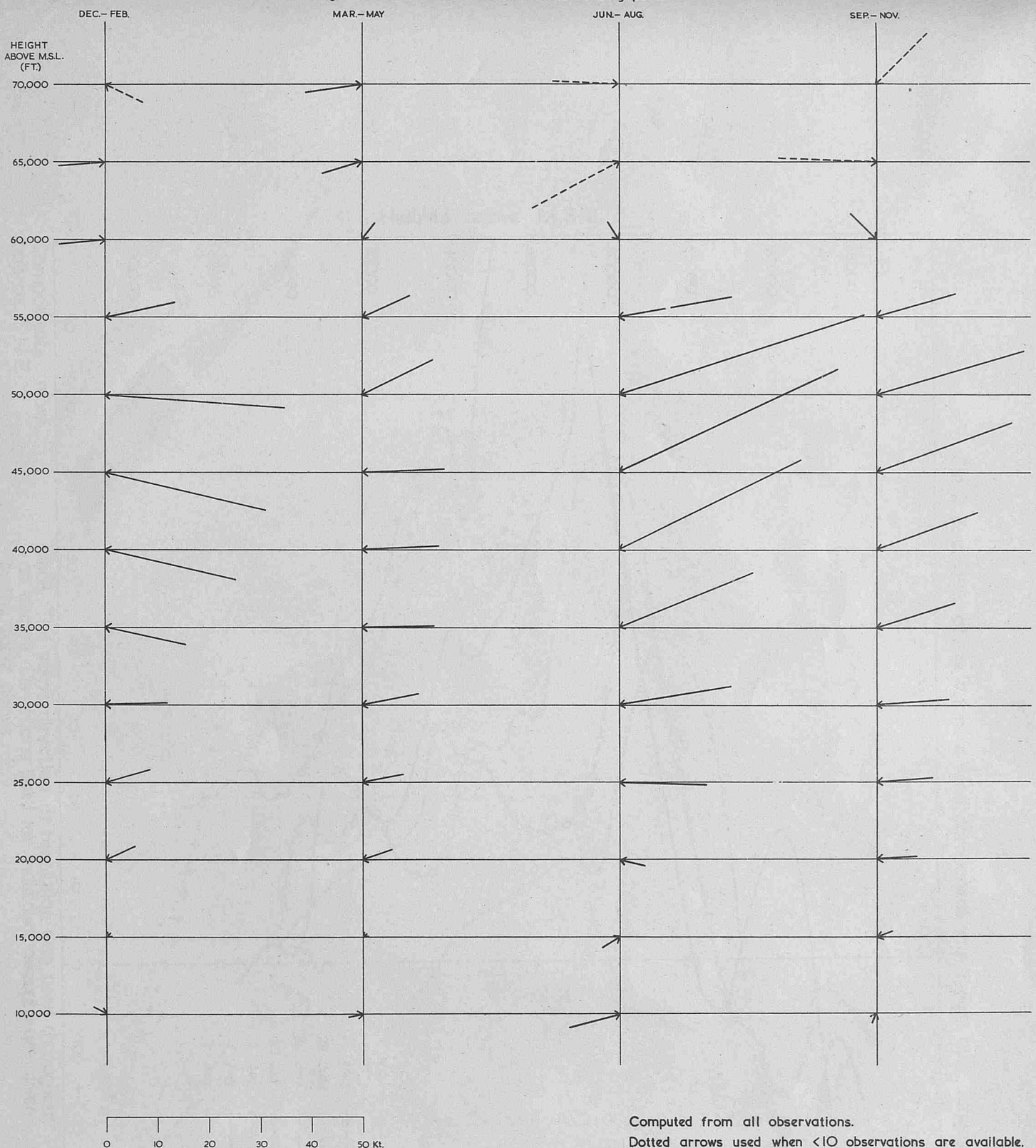
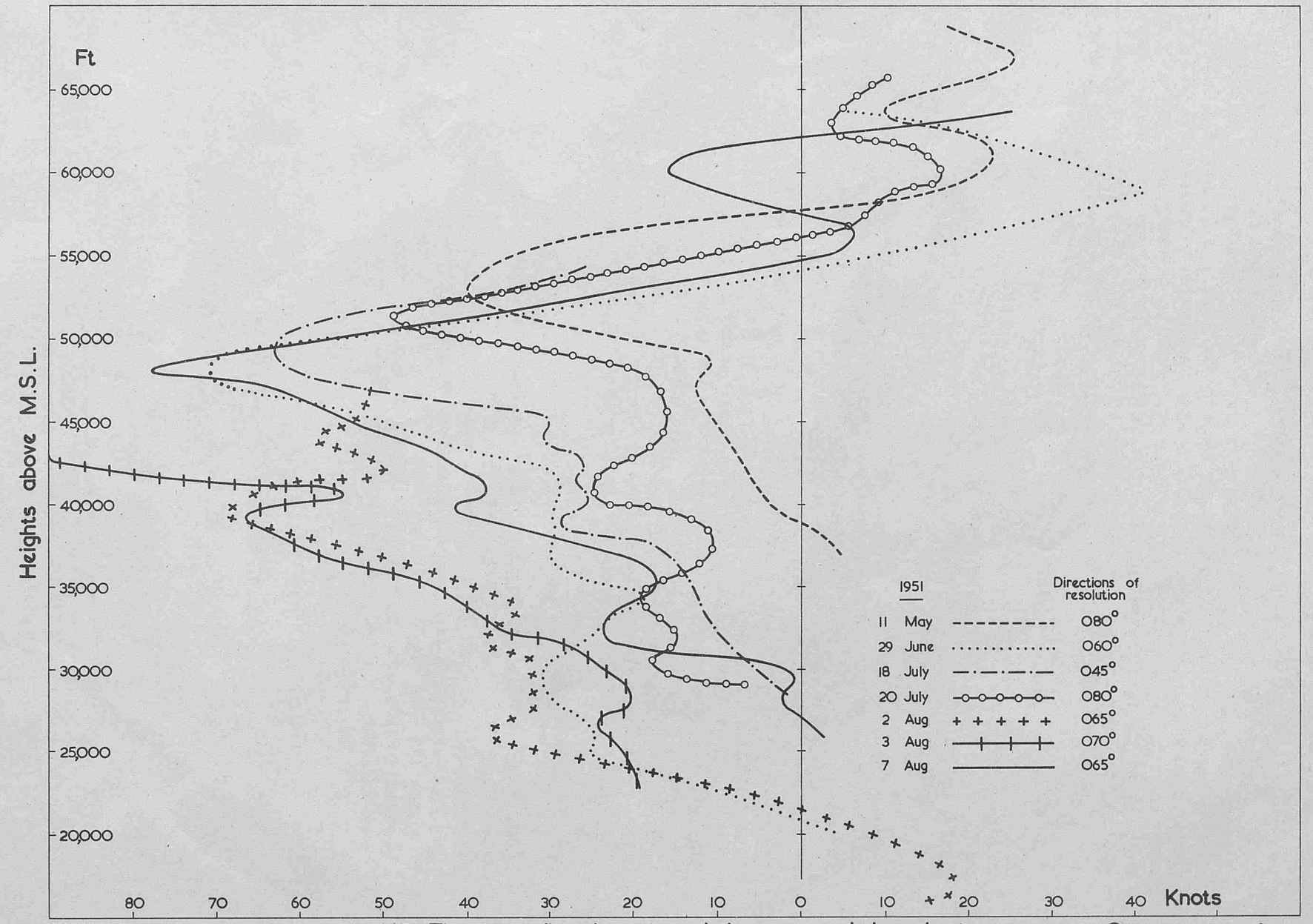
Fig. I. Seasonal vector resultant winds at Singapore.



Computed from all observations.
Dotted arrows used when <10 observations are available.

Fig. 2. Examples of variation of wind component with height over Singapore



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A copy is available in the Library of the Meteorological Office."

M.R.P. 771
S.C.III/144
21st November, 1952

AD 32647

APR 16 1953

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

Turbulence in the upper air, as shown by radar-wind and radio-sonde measurements

by
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AIR WEATHER SERVICE
TECHNICAL REFERENCE FILES

Summary

Observations of the fluctuations, minute by minute of wind components and temperature in a radio-sonde and radar-wind ascent up to 30 km. were analysed. After allowing for the effects of random errors a standard deviation of about 5 knots for the eddy velocity, with a mean wind of about 25 knots, was obtained. The standard deviation of the temperature fluctuations was 0.85°F. in the troposphere but decreased to about one third of this in the upper stratosphere. Rough estimates of the vertical eddy fluxes of heat and momentum and of the coefficients of eddy diffusion were derived from the data.

1. Introduction

In an upper wind sounding the plotted positions of a balloon observed at equal intervals of time fluctuate about the smooth track which the balloon would follow if it were moving with the mean velocity of the wind. There are similar variations in the rate of ascent. The fluctuations are due partly to turbulence and partly to errors of observation. If we know the magnitude of the latter we can obtain some information about the turbulent fluctuations. D.N. Harrison (1948a) has given detailed information about the errors of observation of radar-wind equipment which enable this to be done.

Temperature observations by radio-sonde can be treated similarly since Harrison (1948b) has also obtained estimates of observational error in the temperature readings. The turbulent fluctuations of temperature, however, are smoothed to some extent by the lag of the temperature element.

It seemed worth while applying Harrison's estimates of the errors of observation to one of a series of high altitude ascents to see whether it would yield any useful information on the scale of turbulence at levels up to 100,000 feet.

2. Reduction of the Observations

The sounding chosen for this investigation was one made from Downham Market (52°36'N, 00°24'E) on 29th June, 1949 at 22 hr. G.M.T. and described in a previous paper (Scrase 1949). The synoptic conditions at the time were anticyclonic, there being a high pressure area over Scotland. The smoothed measurements of wind speed, direction and temperature plotted against height are reproduced in Fig.1 from which it will be seen that there was considerable wind shear above and below the tropopause. Radar observations of slant range R, azimuth A and elevation E of the balloon were taken at minute intervals simultaneously with temperature readings from the radio-sonde.

(a) Wind measurements

In order to compare the components of eddy velocity the distances the balloon travelled between one minute and the next along and perpendicular

/(horizontally

(horizontally and vertically) to the bearing of the line of sight were first computed from the radar readings thus:

$$x = R_{n+1}\cos E_{n+1} - R_n \cos E_n$$

$$y = R_{n+1}\cos E_{n+1}(A_{n+1} - A_n)$$

$$z = R_{n+1}\sin E_{n+1} - R_n \sin E_n$$

where the suffixes n and n+1 indicate observations at consecutive minutes. These displacements are plotted against time (and height) in Fig.2 and their deviations from the smooth curves drawn through them represent the observed down wind, crosswind and vertical components of eddy velocity u_o' , v_o' , w_o' for one-minute intervals. Thus:

$$u_o' = (x - \bar{x})/t \; ; \quad v_o' = (y - \bar{y})/t \; ; \quad w_o' = (z - \bar{z})/t$$

where $\bar{x}$, $\bar{y}$ and $\bar{z}$ are the smoothed (mean) values and t is the time interval. The deviations themselves are plotted in Fig.3 and since the long period drifts that are evident in Fig.2 are eliminated the minute to minute fluctuations are more readily comparable.

The standard deviations of the observed fluctuations were computed for each of the three parts of the sounding covered in the periods 3 to 28, 29 to 55 and 56 to 80 minutes from the start and representing the troposphere, the lower stratosphere and the upper stratosphere. The values, each of which was derived from 27 observations, are given in Table I. We now require to know to what extent these values are due to errors of observation.

Harrison's (1948a) investigation of the random errors of radar readings with G.L.III equipment showed the R.M.S. values of the errors to be:

Error in slant range, δR , = 16 yards
Errors in azimuth and elevation δA and δE = 0.10^o = 0.0017 radian.
Using these figures the random errors in the components of eddy velocity derived from the readings can be computed from the expressions:

$$(\delta u)^2 = 2\left[(\delta R)^2 \cos^2 E + R^2 \sin^2 E\,(\delta E)^2\right]/t^2$$

$$(\delta v)^2 = 2\,R^2 \cos^2 E\,(\delta A)^2/t^2$$

$$(\delta w)^2 = 2\left[(\delta R)^2 \sin^2 E + R^2 \cos^2 E\,(\delta E)^2\right]/t^2 \; .$$

These errors were computed for the positions of the balloon at every fifth minute and R.M.S. values for the three parts of the sounding obtained. They are given in Table I together with the standard deviations adjusted by means of the relationship:

$$\overline{(u')^2} = \overline{(u_o')^2} - \overline{(\delta u)^2}$$

The table also includes values of the resultant vector deviation given by:

$$\overline{r'^2} = \overline{u'^2} + \overline{v'^2} + \overline{w'^2}$$

A further question affecting the observed fluctuations is how much does the inertia of the balloon reduce its response to a change in air speed. It

/can

can be calculated that a balloon of the size used (2 kgm.) and with the load carried (also 2 kgm.) should acquire 90 per cent of the velocity of a sudden gust in a few seconds and it can be expected to follow fluctuations having a period of the order of a minute without appreciable reduction of amplitude.

(b) Temperature measurements

The temperature observations, at the minute intervals, were obtained initially as radio-sonde frequencies in cycles per second. A change of frequency of 1 c/s corresponds roughly to a change of temperature of 1°F. The differences between successive readings are plotted in Fig.2. In addition to the fluctuations due to turbulence and to random errors of temperature measurements they include the effect of variations in the rate of ascent; with a vertical gradient of temperature these produce corresponding variations in the temperature differences observed at fixed time intervals. The deviations from the mean curve which are plotted in Fig.3 are, however, corrected for these variations (the magnitude of their effect depends on the temperature gradient). The standard deviations for the three parts of the sounding, when divided by $\sqrt{2}$ and multiplied by the radio-sonde calibration factor, represent the observed temperature fluctuations. They are given in Table I. Harrison's (1948b) investigation of radio-sonde errors gives the R.M.S. random errors of a temperature reading as 0.40°F. We have also to take account of the random error in the measurement of the rate of ascent which, with the temperature gradients observed, cause a random error in the temperature deviations derived from the successive differences, of 0.25°F. in the troposphere and 0.11°F. in the stratosphere. The standard deviations corrected for the combined random error are also given in Table I.

The temperature observations are also subject to errors due to the lag of the bimetal element. For the rate of ascent used the lag coefficient at ground level is 7 seconds and as it probably varies inversely as the square root of the air density the mean coefficients for the three parts of the sounding are approximately 10, 20 and 40 seconds. The effect of lag on the measurement of the temperature fluctuations can be readily allowed for if they are treated as being sinusoidal. There is, in fact, some indication of a rough periodicity of between 3 and 4 minutes in both the temperature and the wind observations. The effect of a lag coefficient λ is to reduce the true amplitude A of a sinusoidal variation of period T to $A/\sqrt{1 + 4\pi^2\lambda^2/T^2}$.

The standard deviations have been corrected on this basis, using the values of λ given above and taking T to be 200 sec; the corrected values are included in Table I.

3. Aerodynamic effects

It will be seen from Fig.2 that the smoothed curve for the rate of ascent of the balloon shows the usual decrease with height after the tropopause is reached. This is generally attributed to the effect of the change in lapse rate on the difference in temperature between the gas in the balloon and the air outside. In the present ascent and similar high altitude soundings with 2 kgm. balloons the decrease in rate of ascent ceases sharply at about 18 km. (60,000 ft.). This is almost certainly due to a change in the drag coefficient* which is associated with the decrease in Reynold's number* as the balloon ascends. In Fig.4 computed values of the drag coefficient are plotted against the logarithm of Reynold's number for comparison with corresponding data quoted by Goldstein (1938) for solid spheres in wind tunnels and dropping in free air. It might be expected that in the steeply changing part of the drag coefficient curve there would be variations in the drag, e.g. due to changes in the shape of the balloon, which would result in increased fluctuations of the balloon speed. The observed fluctuations are, however, not noticeably larger in this part of the ascent than in the parts where the drag coefficient is not changing so much. It seems reasonable to assume, therefore, that the

/fluctuations

* The drag coefficient is $C_o = 8Lg/\rho W^2\pi d^2$ and Reynold's number $Re = W\rho d/\mu$ where L is the free lift, W the rate of ascent and d the diameter of the balloon, and ρ is the density and μ the viscosity of the air.

fluctuations that remain after the observational errors are eliminated are the effect of turbulence in the atmosphere and not turbulence set up by the air flow round the balloon.

4. Eddy velocities and temperature fluctuations

It will be seen from Table I that in all cases the observed standard deviations of wind components exceeded the random errors computed from Harrison's data but the differences are not very large except for the downwind component at all heights and for the crosswind component in the troposphere. The fluctuations in the vertical component were smaller than the horizontal fluctuations at all heights. In the troposphere the crosswind fluctuations were the largest but in the stratosphere the downwind deviations were largest. The vector standard deviation of 5.2 knots for the whole sounding was about 20 per cent of the mean wind speed. This ratio is practically the same as was found by Giblett and others (1932) for the horizontal component of surface wind at Cardington, using 5 second means. Other data for comparison include a series of observations on smoke puffs at heights varying between 2,000 and 27,000 feet from which Durst (1948) obtained vector standard deviations (in the horizontal plane) ranging from 0.5 to 3.5 knots, the average being 1.5 knots, for an average wind speed of 18 knots; the downwind and crosswind components were practically the same size. Durst and Gilbert (1950), from observations of constant height balloons drifting at about 20,000 feet in winds averaging about 50 knots, obtained fluctuations of horizontal wind of the order of 3 or 4 knots from one half-hour to another. Voss (1950) observed fluctuations in rates of ascent in radio-sonde observations which were greater than observational errors and confirmed their magnitude by using radio-sonde heights computed from simultaneous pressure and temperature measurements. He obtained, in clear air up to 20 km. (65,000 ft.), maximum vertical velocities of about 4 knots, the periods of the fluctuations averaging about 3 minutes.

The periods of the wind fluctuations observed in the Downham Market sounding were practically the same for each component and were roughly 3 to 4 minutes. Oscillations of about this period, in which the up and down currents were about 5 - 10 knots, were recently reported by an aircraft flying at 8,500 feet near the West of Scotland (Turner 1952). In the Downham Market sounding the period of oscillation would correspond to eddies having horizontal dimensions of about 10,000 feet and vertical dimensions of 5,000 feet. In Durst's (1948) smoke puff observations the periods of the fluctuations were 5 to 15 minutes and the eddies extended horizontally 2 to 10 miles and vertically 5,000 feet or more. Voss's observations indicated eddy dimensions of about 6,000 feet horizontally and 2,500 feet vertically. The Cardington measurements of surface wind showed the major eddies to be elongated in the downwind direction and to be about 4,000 feet long and 600 feet wide but there were other groups of major eddies, in certain conditions, which had dimensions in the direction of the wind up to 10, and even 20, miles.

The maximum minute to minute velocities measured in the Downham Market sounding were about 1,500 feet/min. for the horizontal components and about 800 ft./min. for the vertical component; they all occurred in the layer between 30,000 and 60,000 feet. The maximum minute to minute accelerations were 0.02 g. for the horizontal and 0.01 g. for the vertical components.

Turning now to the temperature fluctuations it will be seen from Table I that in the upper stratosphere the standard deviation is not much larger than the random error but in the troposphere it is roughly twice as large as the error and after allowing for this and for lag a value of 0.85°F. is obtained. The corresponding value for the lower stratosphere is 0.64°F. Fluctuations of this order of magnitude have been reported by Frith (1950) and by Grant (1951) from aircraft observations in the troposphere and in the lower stratosphere. The period of the temperature fluctuations observed at Downham Market was about the same as for the wind fluctuations.

5. Estimates of momentum and heat fluxes and the coefficients of eddy diffusion

The simultaneous observations of u', v', w' and T' together with measurements of the vertical gradients of wind and temperature shown in

/Fig.1

Fig.1 should enable rough estimates to be obtained of the vertical eddy fluxes of linear momentum (i.e. eddy shearing stresses) and heat and the coefficients of eddy diffusion for momentum and heat, using the formulae:

Eddy flux of momentum in direction of the wind $\approx \overline{\rho}\ \overline{u'w'}$

Eddy flux of momentum across wind $\approx \overline{\rho}\ \overline{v'w'}$

Eddy flux of heat $\approx c_p\ \overline{\rho}\ \overline{T'w'}$

Coefficient of eddy viscosity in direction of wind $K_{Mu} = -\overline{u'w'} / \frac{d\overline{u}}{dz}$

Coefficient of eddy viscosity across wind $K_{Mv} = -\overline{v'w'} / \frac{d\overline{v}}{dz}$

Coefficient of eddy conductivity $K_H = -\overline{T'w'} / \left[\frac{d\overline{T}}{dz} - \left(\frac{d\overline{T}}{dz}\right)_a\right]$

where ρ is the air density, c_p the specific heat at constant pressure and $(dT/dz)_a$ the adiabatic temperature gradient (taken as -1.8×10^{-4} °F cm^{-1} for the subsequent computations).

Since the observed fluctuations do not greatly exceed the random errors of observation, especially in the higher levels, it is doubtful whether estimates derived from the present observation can be of very great value. Any correlations that exist between the turbulent fluctuations may be masked by correlations between the observational errors. However, we have some knowledge of these latter correlations and, in theory, can allow for them by relationships of the type

$$\overline{u_o' w_o'} = \overline{u'w'} + \overline{\delta u'\ \delta w'}$$

if we know or assume that there is no correlation between the random errors and the true fluctuations.

We have no direct measurements of $\overline{\delta u'\ \delta w'}$ etc. but can derive the values from the standard errors of the observations of range, azimuth and elevation provided we know how much correlation, if any, there is between them. The data used by Harrison (1948a) were examined and no significant correlations between $\delta R'$ and $\delta A'$ and between $\delta R'$ and $\delta E'$ were found, but $\delta A'$ and $\delta E'$ gave a correlation coefficient of about -0.3. The mean values of the products required could then be computed from:

$$\overline{\delta u'\ \delta v'} = +2\ \overline{R^2 \sin E \cos E}\ 0.3\ \delta E\ \delta A$$

$$\overline{\delta u'\ \delta w'} = -2\ \overline{R^2 \sin E \cos E}\ (\delta E)^2 + 2\ \overline{\sin E \cos E}\ (\delta R)^2$$

$$\overline{\delta v'\ \delta w'} = -2\ \overline{R^2 \cos^2 E}\ 0.3\ \delta E\ \delta A$$

in which δE, δA, δR are R.M.S. values.

There is no reason to expect any correlation between temperature errors and δw other than that due to errors in measuring the height intervals

/(or

(or rate of ascent) in which case it can be shown that:

$$\overline{T_o'w_o'} = \overline{T'w'} - \frac{\sqrt{2}}{t}\left(\frac{\overline{dT}}{dz}\right)(\overline{\delta z})^2 = \overline{T'w'} - \frac{t}{\sqrt{2}}\left(\frac{\overline{dT}}{dz}\right)(\overline{\delta w})^2.$$

Any correlation between T' and w' , however, is also affected, especially at the higher levels, by the thermometer lag; but again this can be allowed for if we assume, as before, that the fluctuations are sinusoidal. In this case, however, phase as well as amplitude distortion is involved and $\overline{T'w'}$ is reduced by a factor $1/(1+4\pi^2\lambda^2/T^2)$.

Estimates of the eddy fluxes and the coefficients of eddy diffusion derived from the lower parts of Downham Market sounding (troposphere and lower stratosphere) and corrected, in the manner indicated above, for the effects of random errors and lag are given in Table II. It was not considered worth while including estimates for the upper stratosphere part of the sounding since there the effects of random errors overwhelmed any correlations that there may have been between the natural fluctuations. The results for the lower stratosphere also are of doubtful value and should not be regarded as more significant than the order of magnitude. In the troposphere, however, where the observational errors are smaller, the results are more significant and are perhaps worth quoting since there is a scarcity of such information for the free atmosphere.

The heat flux of 10 mw. cm.$^{-2}$ found for the troposphere is of the same order of magnitude as for small scale turbulence near the ground. Thus, Swinbank (1951) found values of about -1 and 5 mw. cm.$^{-2}$ depending on the lapse rate, at a height of 1.5 m. and Lander and Robinson (1952) obtained values between -4 and 7 mw. cm.$^{-2}$ at 25 cm. above the ground. For comparison with the momentum flux $\rho\,\overline{u'w'}$ of -1.7 dynes cm.$^{-2}$ found in the present investigation, we may quote the values of -3.6 and -0.8 dynes cm.$^{-2}$ measured at heights of 19 m. and 1.5 m. respectively by the author (1930) who also found $\rho\,\overline{v'w'}$ at 19 m. to be -1.5 which may be compared with -5.6 dynes cm.$^{-2}$ for the troposphere in the present investigation.

The estimates of the coefficients of eddy diffusion for the troposphere in the Downham Market sounding are reasonably consistent with the relationship found by Richardson (1926) between the diffusion coefficient and the scale of turbulence:

$$K = 0.2\, l^{4/3},$$

Richardson found this applied to five groups of observations ranging in scale from $l = 1.5 \times 10^3$ cm. to $l = 5 \times 10^6$ cm. In our case l lies somewhere between the vertical distance travelled by the balloon in one minute, i.e. about 4×10^4 cm., and the distance covered in 27 minutes, i.e. about 10^6 cm. The rough periodicity of the fluctuations of about 200 sec. suggests a value for l of 10^5 cm., for which Richardson's relationship would give K about 10^6.

It will be seen that the troposphere value of the crosswind eddy viscosity is nearly five times the value in the direction of the wind. Richardson (1920) found ratios of this order of magnitude from Lindenberg pilot balloon observation at heights of between 0.2 and 0.5 km.

The troposphere value of the eddy conductivity was 0.3 times that of the eddy viscosity (in the direction of the wind). This is about half the ratio (0.65) deduced by Petterssen and Swinbank (1947) for large scale turbulence in the atmosphere.

/It

It will be noticed that in the Downham Market ascent the eddy flow of heat in the troposphere was directed upwards and K_H was negative although the potential temperature increased with height. This could be taken as supporting the view of Ertel (1942) who questioned the assumption in the classical theory of eddy diffusion that the eddy flux is in the direction of the gradient of potential temperature and is therefore normally downwards. Ertel concluded that when buoyancy effects are taken into account the flux is more likely to be determined by the gradient of actual temperature. Priestley and Swinbank (1947) put forward similar ideas and suggested a modified form for the eddy flux having a component due to buoyancy forces which is always directed upwards and thus is generally opposed to the downward transfer by mechanical turbulence. In lapse rates less than the adiabatic the coefficient of eddy diffusion can then assume both positive and negative values. The theory is capable of extension to some other properties such as momentum and might therefore also explain the negative eddy viscosity K_{Mu} in the lower stratosphere of the Downham Market ascent where the wind velocity gradient was negative. It also explains why the diffusion coefficients for different properties do not necessarily have the same values.

6. Future work

The present sounding technique, with eye readings at minute intervals, is not designed for the measurement of turbulent fluctuations in the free atmosphere but this investigation indicates that with some improvements to reduce the observational errors useful measurements should be obtainable. The chief need is for a higher degree of angular resolution, especially in elevation, in the wind finding equipment. Automatic tracking of the balloon should facilitate this and when combined with automatic recording of the data it should enable readings to be obtained at shorter time intervals. A considerable increase in the rate of response of the temperature element is desirable and analysis of the data would be facilitated by automatic computing of the wind components. Such improvements are, in fact, being incorporated in the radar-sonde theodolite system now under development (Jones, Hooper and Alder 1951).

Acknowledgment

I am glad to acknowledge the assistance of Miss D. Chapman and Mrs. J. Seaton in the laborious task of tabulation and computation which this investigation has involved.

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Table I

Standard deviations of wind component and temperature fluctuations from minute to minute, at Downham Market, at 20.00h. G.M.T. on 29th June, 1949

Time from start	min.	3-28	29-54	55-80	3-80
Height	km.	1-12	12-21	21-31	1-31
Mean wind speed	knots	29	23	20	24
Mean rate of ascent	knots	13	12	11	12
Mean wind direction		340°	020°	090°	

		Knots			
Along wind					
Observed	u_o'	1.8	5.1	4.6	4.1
Random error	δu	0.7	1.4	2.1	1.5
Corrected	u'	1.7	4.9	4.1	3.8
Crosswind					
Observed	v_o'	4.0	3.6	3.7	3.8
Random error	δv	0.9	2.5	2.7	2.2
Corrected	v'	3.9	2.6	2.4	3.1
Vertical					
Observed	w_o'	1.5	3.0	3.6	2.8
Random error	δw	1.0	2.6	2.8	2.3
Corrected	w'	1.1	1.5	2.2	1.6
Vector Resultant					
$(\overline{u'^2} + \overline{v'^2} + \overline{w'^2})^{1/2}$		4.4	5.7	5.2	5.2

Temperature fluctuations				
		°F.		
Observed	T_o'	0.94	0.67	0.45
Random error *	δT	0.49	0.41	0.41
Corrected for random error		0.81	0.53	0.17
Corrected for random error and lag	T'	0.85	0.64	0.29

* Combines height interval error and temperature error

Table II

Vertical eddy fluxes and coefficients of diffusivity derived from Downham Market sounding of 20.00h. G.M.T. on 29th June, 1949

Height	km.	1-12	12-21
Mean density, $\overline{\rho}$	gm. m^{-3}	700	200
$\overline{du/dz}$	sec.$^{-1}$x10^{-3}	+1.9	-3.3
$\overline{dv/dz}$	sec.$^{-1}$x10^{-3}	+1.4	-0.6
$\overline{dT/dz} - (dT/dz)_a$	°F.cm.$^{-1}$x10^{-4}	+0.7	+2.0
Vertical eddy fluxes			
Momentum parallel to wind, $\overline{\rho}\,\overline{uw}$	dynes cm.$^{-2}$	-1.7	-1
Momentum across wind, $\overline{\rho}\,\overline{vw}$	" "	-5.6	0
Heat, $c_p\overline{\rho}\,\overline{Tw}$	mw. cm.$^{-2}$	+10	-5
Coefficients of eddy diffusion			
Viscosity parallel to wind K_{Mu}	cm.^{2}sec.$^{-1}$x10^5	+13	-14
Viscosity across wind K_{Mv}	"	+57	+1
Heat conductivity, K_H	"	-4	+2

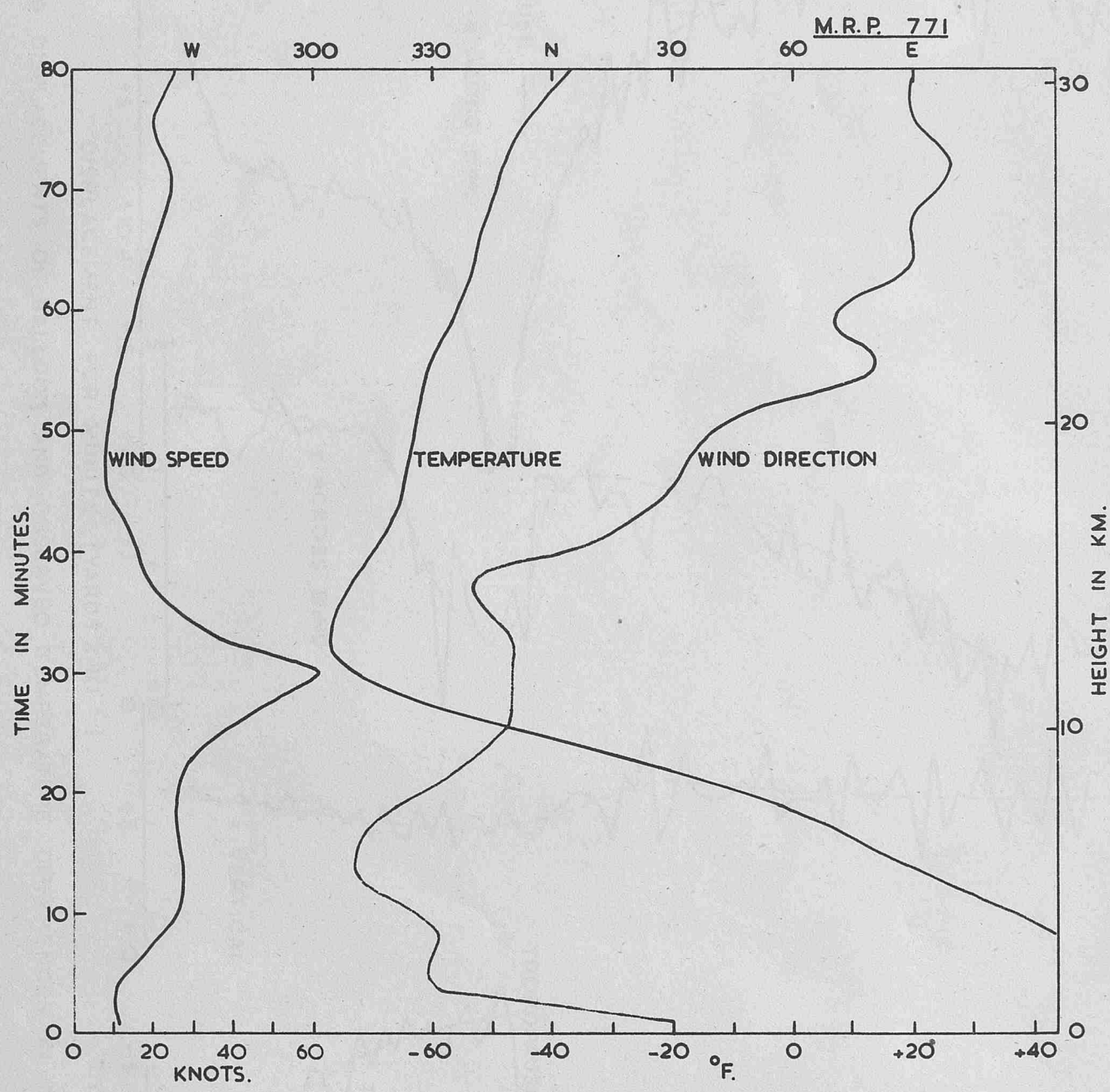

FIG. No. 1. Upper air sounding at Downham Market at 2200hrs. G.M.T. on 29th June 1949.

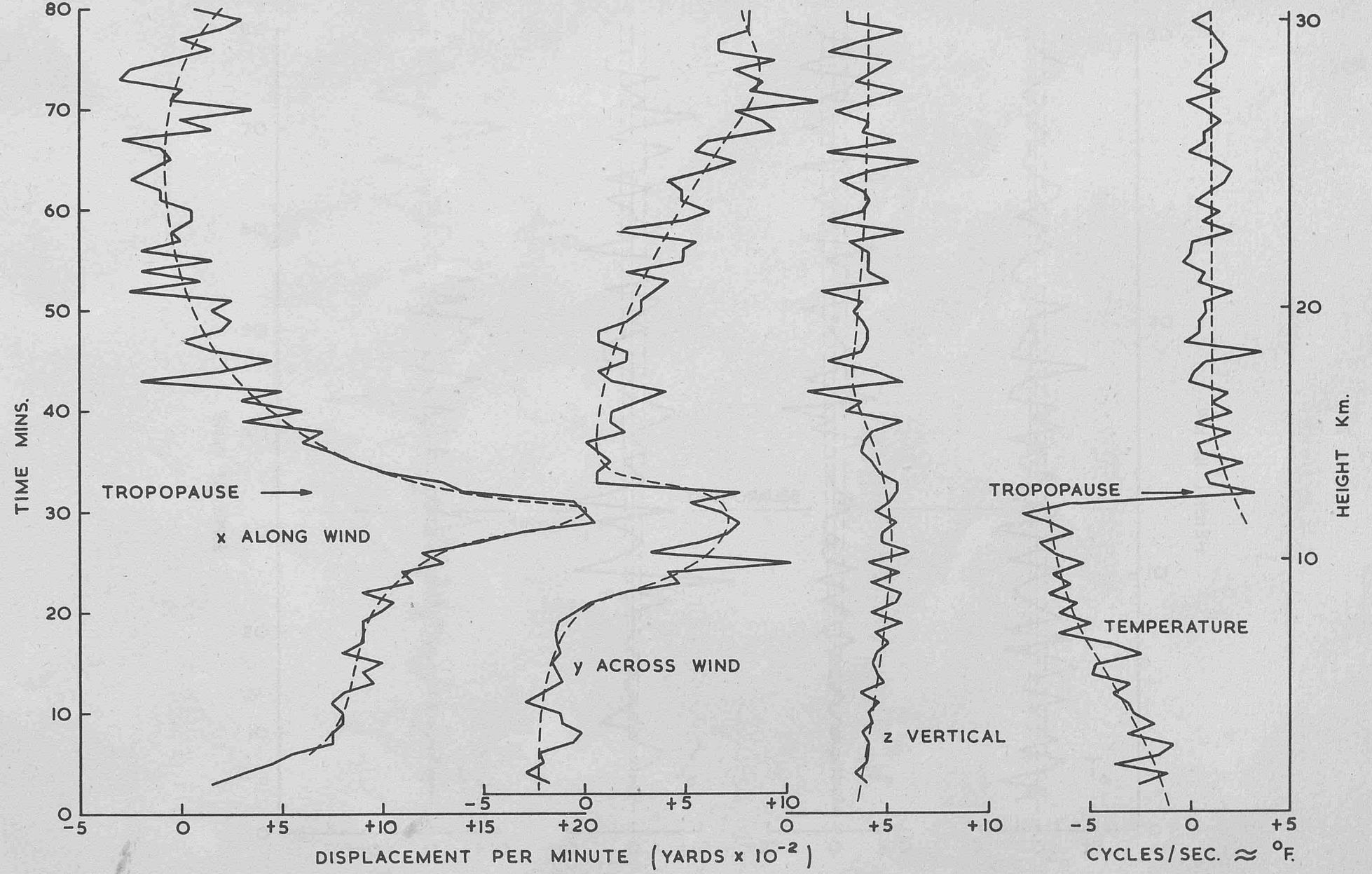

Fig. 2 OBSERVED DISPLACEMENTS OF BALLOON AND OBSERVED TEMPERATURE DIFFERENCES IN ONE MINUTE INTERVALS

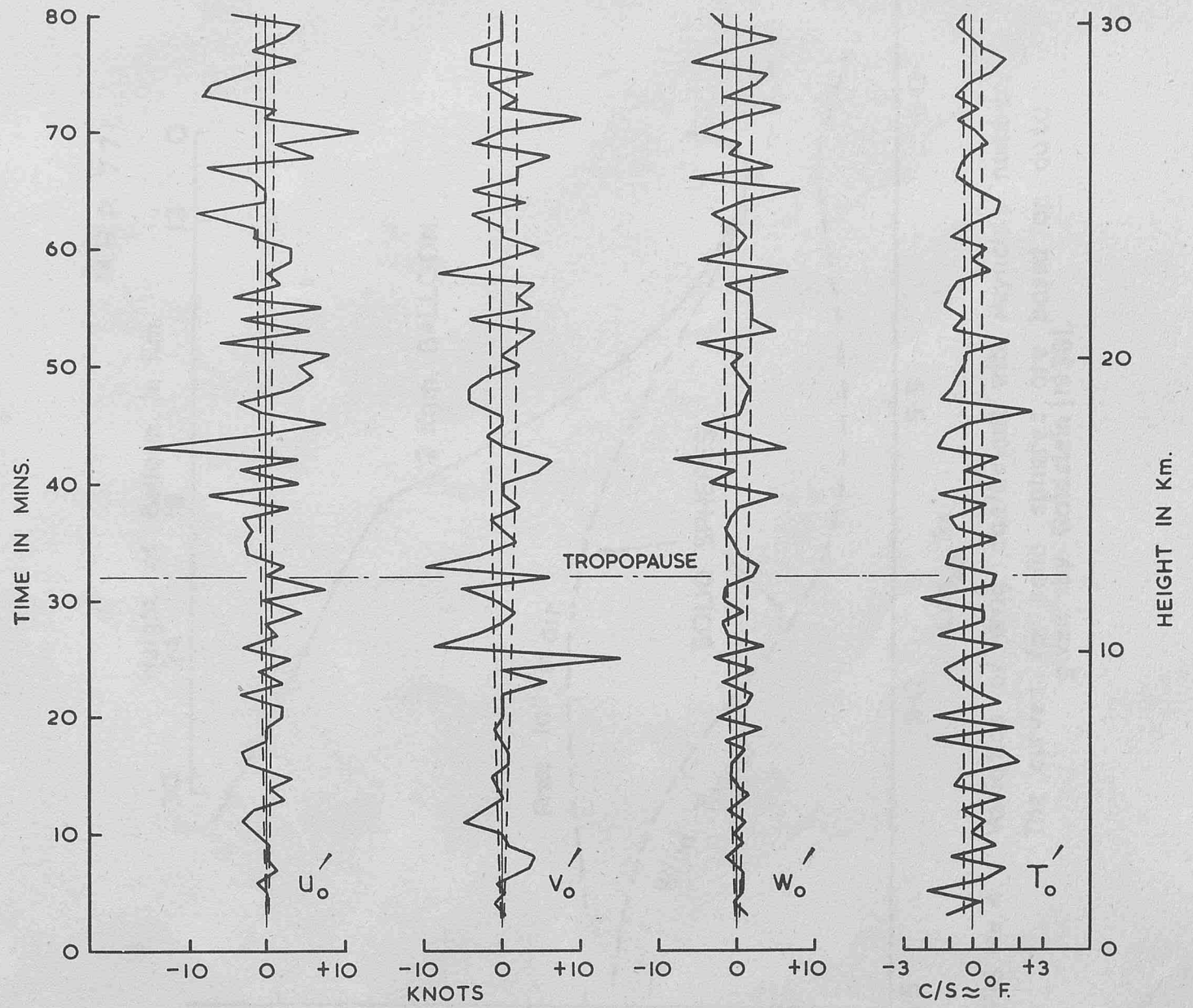

Fig. 3. OBSERVED DEVIATIONS OF VELOCITY COMPONENTS AND TEMPERATURE DIFFERENCES FROM MEAN CURVES (DOTTED LINES INDICATE PROBABLE ERROR OF INDIVIDUAL OBSERVATIONS)

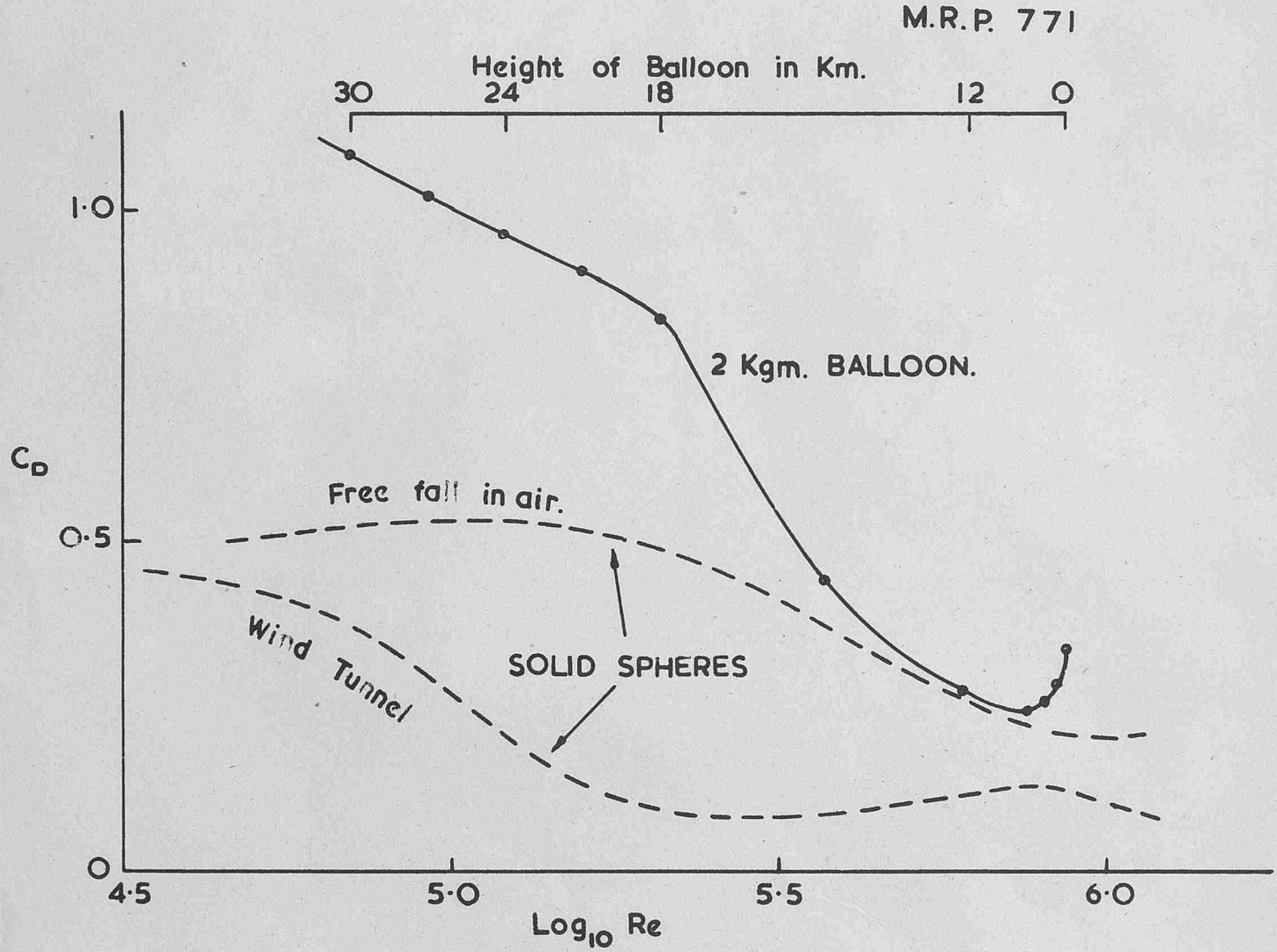

FIG. No. 4. Variation of drag coefficient with Reynolds number. The curves for solid spheres are based on data given by Goldstein (1938).

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No..... by Mr.....
A copy is available in the Library of the Meteorological Office."

M.R.P. 772 (Revised)
S.C.II/131

24th November, 1952.

AD 32646

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

The Profile of a Jet Stream observed 1st September, 1952.

by
G.W. HURST (Meteorological Office)

In M.R.P. 722, the results obtained from a flight across a marked jet stream on 18th January, 1952 were discussed, and a cross-section at approximately 300 mb of the leading edge and the main axis of the stream was derived. A similar flight of the same nature was made on 1st September, 1952 and the results are discussed below; this flight was in many respects fuller than the earlier one.

The meteorological situation which obtained at 0600 GMT on 1st was that a cold front running from South Denmark to Belgium and thence to NW France was moving east at about 30 knots in the north, with a slight southerly drift over NW France. An anticyclone to the south west of the British Isles was gaining in intensity, and subsidence was becoming effective over the south of the country. Winds from a general direction of 260° were reported over the British Isles as a whole at 0900 GMT with a variation in speeds at 300 mb from just over 40 knots in the extreme south (Camborne and Larkhill) to well over 100 knots in South Scotland (Leuchars 120 knots and Aldergrove 105 knots) and to below 60 knots in the extreme north (Stornoway and Lerwick). In view of the expectation of very well broken cloud in the south, and breaks between showers in the north, it seemed worth while to arrange for a cross-sectional flight. Accordingly, F/O's Slaymaker and Naldrett were briefed to fly at 30,000 ft. indicated and altimeter sub-scale setting of 1013 mb (almost exactly 300 mb) on a course of 335° magnetic from Bedford to north of Scotland; it was calculated that Cape Wrath would be overflown. It was arranged that vertical photography should take place using a time interval of 15 seconds.

The flight took place as planned. The aircraft positioned over Bedford at 1045 GMT, and a constant magnetic heading of 335° and height of 30,000 ft. (indicated) was maintained at an indicated air speed of 165 knots (equivalent to true air speed of 272 knots in the south and 266 knots in the extreme north) and the north coast of Scotland was crossed at 1220 GMT at a point about 5 miles SW of Cape Wrath. As far north as Sheffield cloud was insufficient to prevent ground identification of detail for more than an odd print here and there. To the north of this however there were gaps in the series of ground photographs; the only ones which gave rise to more than 15 n.m. occurred west of Hexham (19 n.m.) over the Cheviots (30 n.m.) and just south east of Cape Wrath (17 n.m.). As far as possible, mean speeds were calculated between intervals of five prints (approximately 5 n.m.) but particularly on the occasions mentioned above, this procedure had necessarily to be departed from materially. The standard of flying was for the most part very high, and the only part of the flight for which satisfactory winds could not be deduced was in the extreme north of Scotland from 58°15' - 58°30'N. Careful Gee plots were maintained by the navigator and winds were deduced. The flight was carried through to 59°N., but no detailed analysis has been made north of Cape Wrath.

The profile of the wind speed at flight level is shewn in Figure 1, on which both the wind calculated from photographs (continuous lines) and those from Gee fixes (pecked lines) are plotted. It will be seen that the maximum wind calculated was 112 knots at 55°50'N., and that the breadth of the jet stream axis (considering this as winds within 10 knots of the maximum) was from 55°15' to 55 55'i.e. 40 n.m. It is unfortunate that the major gap in fixes due to cloud cover occurred in just this vicinity, so that the peak

/wind

wind speed of the stream might well have proved to have been about 115, or even 120 knots. The fall off of wind speed on the leading (cold) edge was from 112 knots to 45 knots in a distance of about 75 n.m. (a gradient of 67 knots/75 n.m., compared with the stronger January occasion of 90 knots/80 n.m.). From this point, 75 n.m. north of the edge of the jet stream, wind remained almost steady at 45 knots for some way; there is a break from 58°15' - 58°30'N., and in the extreme north, the wind measured 39 knots. The Gee winds over the whole of this area were rather stronger than 40 knots.

On the warm side of the stream the increase from the south was more uniform and from the start of photography at Bedford to the southern edge of the jet stream axis the increase of wind was 40 to 112 knots i.e. 72 knots in 200 n.m. It will be noted on this diagram that there is very close agreement between the winds calculated from photographic analysis, and those derived from Gee fixes. It is perhaps of interest that the maximum ratio of anticyclonic shear to the Coriolis parameter was .83, so the possibility of dynamic instability does not arise. The cyclonic shear/Coriolis parameter has a maximum value of 1.75.

In view of the stepped nature of the winds profiled in Figure 1, an assessment has been made of the errors involved in wind computation. Errors can arise from three main sources:

(a) Errors in locating the print centre.

(b) Variation of the attitude of the aircraft during flight.

(c) Variation of the heading of the aircraft in flight.

Of these (a) is almost negligible. Prints can be identified and marked on a one inch Ordnance Survey map with an accuracy of about one millimetre (.04 mile) and over a series of five prints the probable error from these sources is 1.2 knots. The error has also been calculated, for reasons explained below, on six and a half prints; this proved to be .9 knots.

Errors arising from (b) are due to yawing, pitching and rolling. Of these, yawing does not affect the centre of the print, and the effect of pitching is primarily on direction, and only secondarily on speed. Rolling however can affect the deduced wind significantly. It is an almost instantaneous phenomenon (a matter of at most a few seconds) and according to the pilot it should not exceed 2°. This extreme error could give rise to a ground displacement of about .20 mile, and a wind speed maximum error on six and a half prints of 12 knots. This has largely been eliminated as correction has been made before taking measurements on any fix which was plainly out of line both with the track and with its immediate neighbours. It is of course possible that some element of real wind variation might exist in such cases, but it is thought much more likely that roll is the culprit. About half the fixes were adjusted slightly, though very few adjustments of more than .1 mile were necessary. It is difficult to say what residual probable error may remain after this correction, but is thought to be less than 2 knots over five prints, and 1.5 knots over six and a half prints.

The most important incorrectable source of error in wind estimation comes from uncertainties of the exact heading of the aircraft (c). The flight was done with meticulous care, but the pilot pointed out that the D.I. gyro compass precesses up to 3° every fifteen minutes or so, and that the D.R.C. compass might shew readings up to an error of about 2. He therefore considers that the angle between flight and course would very seldom be as much as 2°, and in most cases deviation would not be greater than 1°. Accepting 2° as the definition of three times the probable error in this case, the possible error in wind speed rising from this source would be 3.2 knots.

The combined probable error over five prints from the above assumptions is 4.0 knots and over six and a half prints 3.6 knots.

/An

An analysis was made of the profile in Figure 1. It was found that on the warm side of the stream the increase in speed in the print range 60 - 185 (about 135 n.m.) was almost linear and a standard deviation was calculated from a linear assumption, and the actual departures from it. The value proved 5.2 knots, giving a probable error of 3.5 knots; the mean print interval was six and a half. A similar calculation was made on the cold side and a probable error of 3.1 knots emerged over (fortuitously) the same print interval. As these values are both lower than the probable error of 3.7 knots in the derived wind values it can be concluded that the winds obtained are not inconsistent with a smooth curve, and that possible errors due to wind derivation have in fact been somewhat over-emphasised.

Figure 2 shews the trajectory of the flight as a whole from Bedford to the north west of Scotland together with vectors representing the winds derived from the photographs. A point which will be immediately noticed is the variation in direction of the wind. In the extreme south, the general direction is 280°, and there is a gradual backing of the wind as the jet stream is approached; the axis of the stream is itself from 245°, and to the north of the axis the direction veers to almost due west. This is related to the contour pattern at 300 mb wherein the track lies roughly along a trough, and flight is oscillating about the trough line.

/Figure

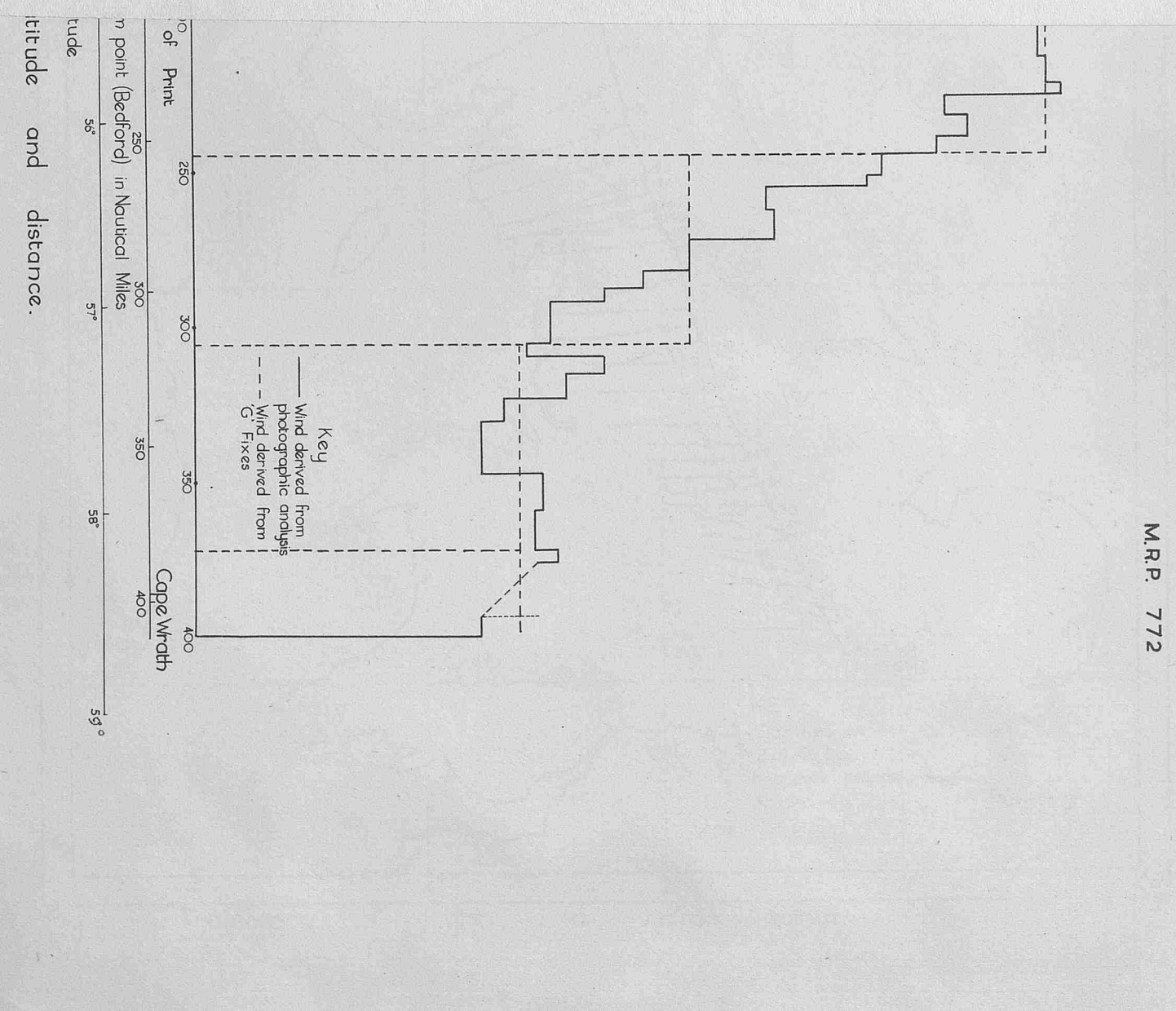

Figure 3 has been drawn to shew the cross section of the flight area at 1500 GMT, the nearest time possible to the time of flight. As many radiosonde and wind ascents as possible have been used by means of projection across the track. Difficulty was experienced in deciding the best angle of projection, and in the end a projection of a standard 260° (a mean wind) was used for all stations; this was almost perpendicular to the mean line of flight. There was a strong thermal field over the area of flight and the double tropopause at Aldergrove brings out the discontinuous nature of the tropopause on this occasion. The wind field was drawn in terms of the actual speeds, and not of components perpendicular to the track. A compromise had to be made on this point - it was clearly more desirable to use components than actual winds, but at levels between 600 and 300 millibars a trough with an associated 30° wind change lay between Aldergrove and the flight line, and it was felt that resolution perpendicular to the track was hardly admissable in such a case. Both actual and resolved winds are indicated on the diagram. One or two curious points in the wind structure arose - the maximum Leuchars wind in the zone 250 - 200 millibars, well above the tropopause, and the secondary maximum wind speed at Aldergrove at 450 millibars are both rather anomalous. Figure 3 is diagrammatic only, as in view of the distance of some of the stations from the track, and the projection difficulties due to the trough near the line of flight, a quantitative agreement between the thermal and the wind field cannot fairly be evolved. In particular it will be noticed that the anticyclonic shear appears as strong as the cyclonic shear, though at the level of flight at least it was only about half as great.

a one inch Ordnance Survey map with an accuracy of about one millimetre (.04 mile) and over a series of five prints the probable error from these sources is 1.2 knots. The error has also been calculated, for reasons explained below, on six and a half prints; this proved to be .9 knots.

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The combined probable error over five prints from the above assumptions is 4.0 knots and over six and a half prints 3.6 knots.

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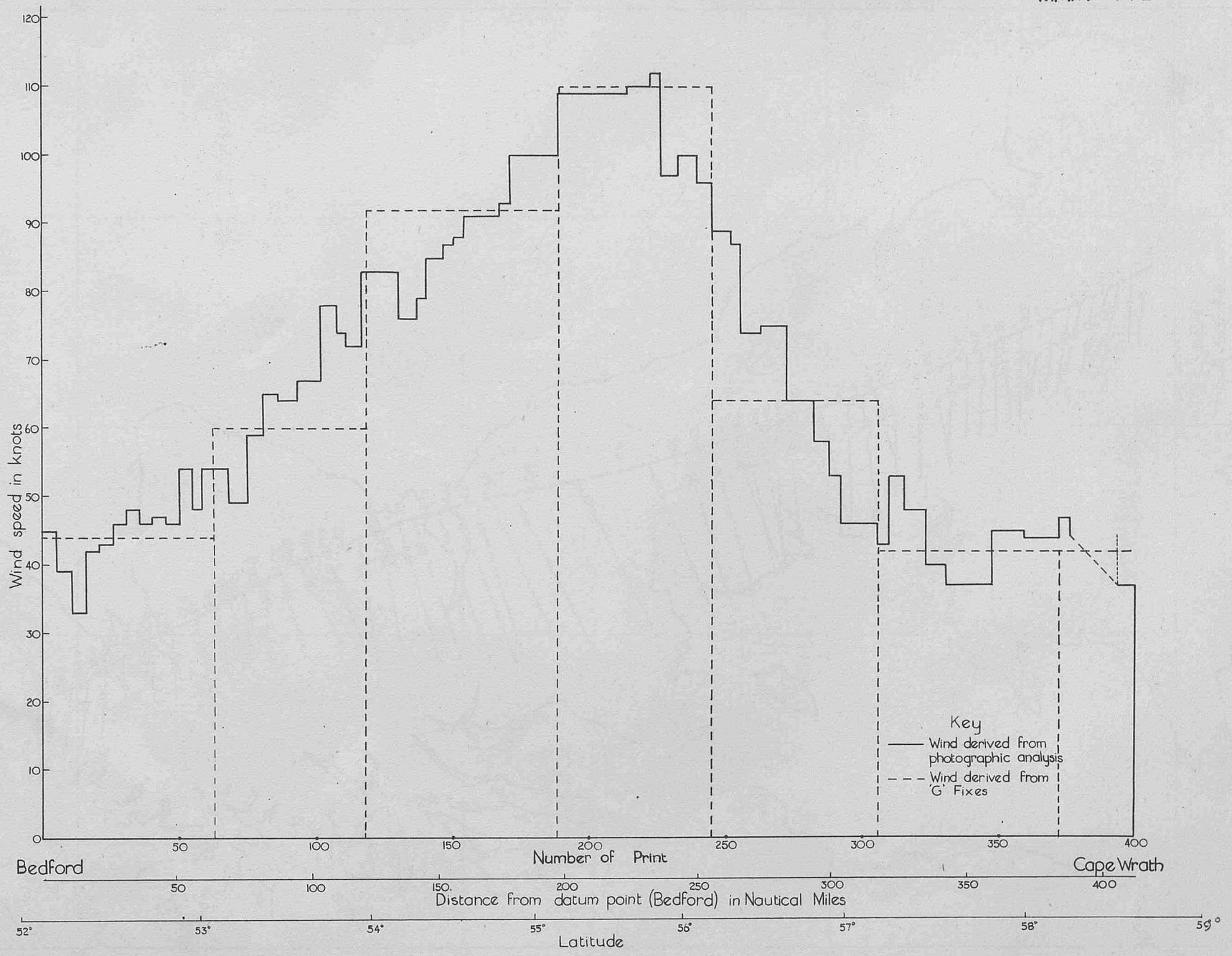

Fig. 1 Variation of wind strength with latitude and distance.

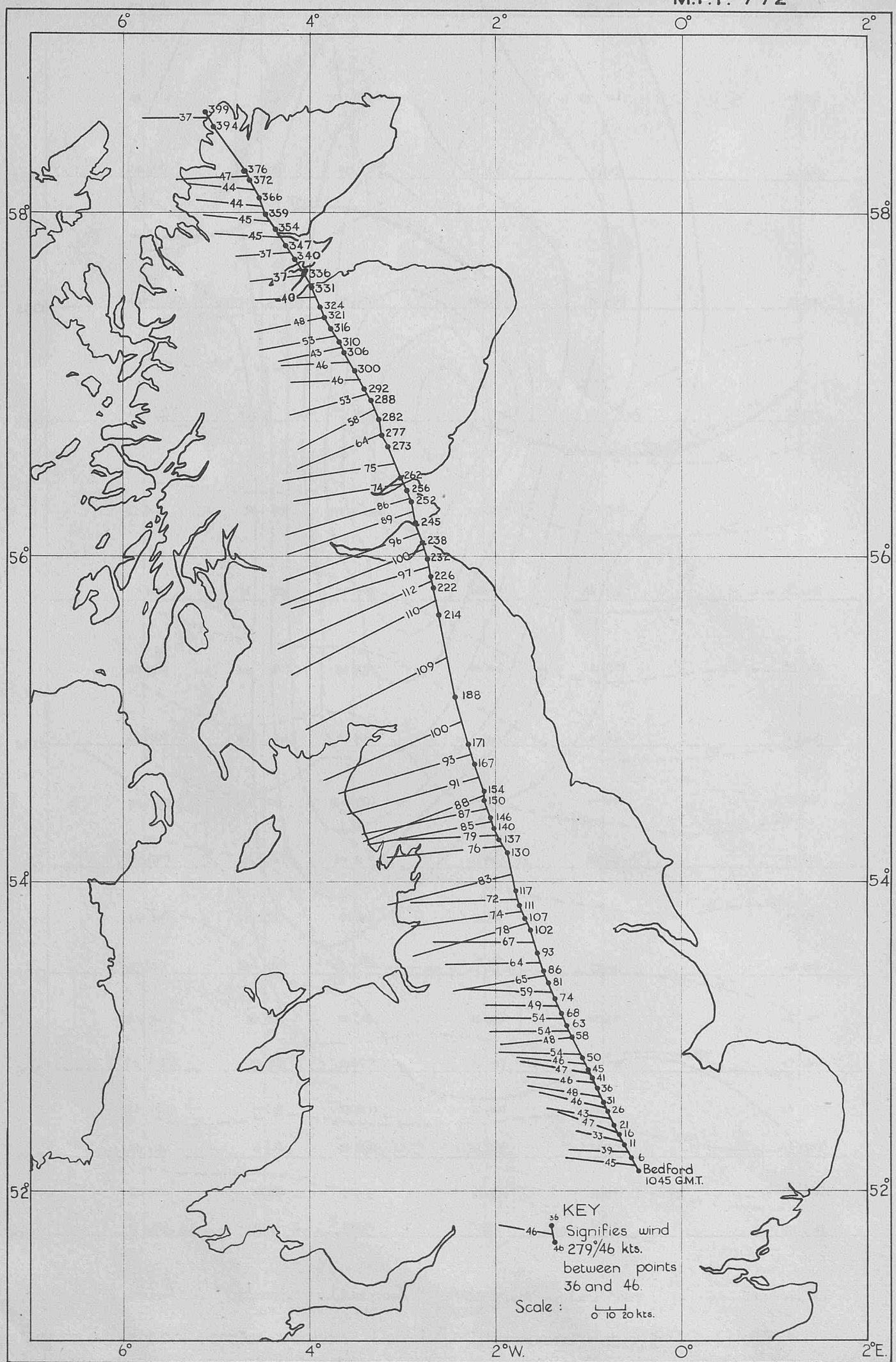

Fig. 2. Trajectory of flight and vector winds.

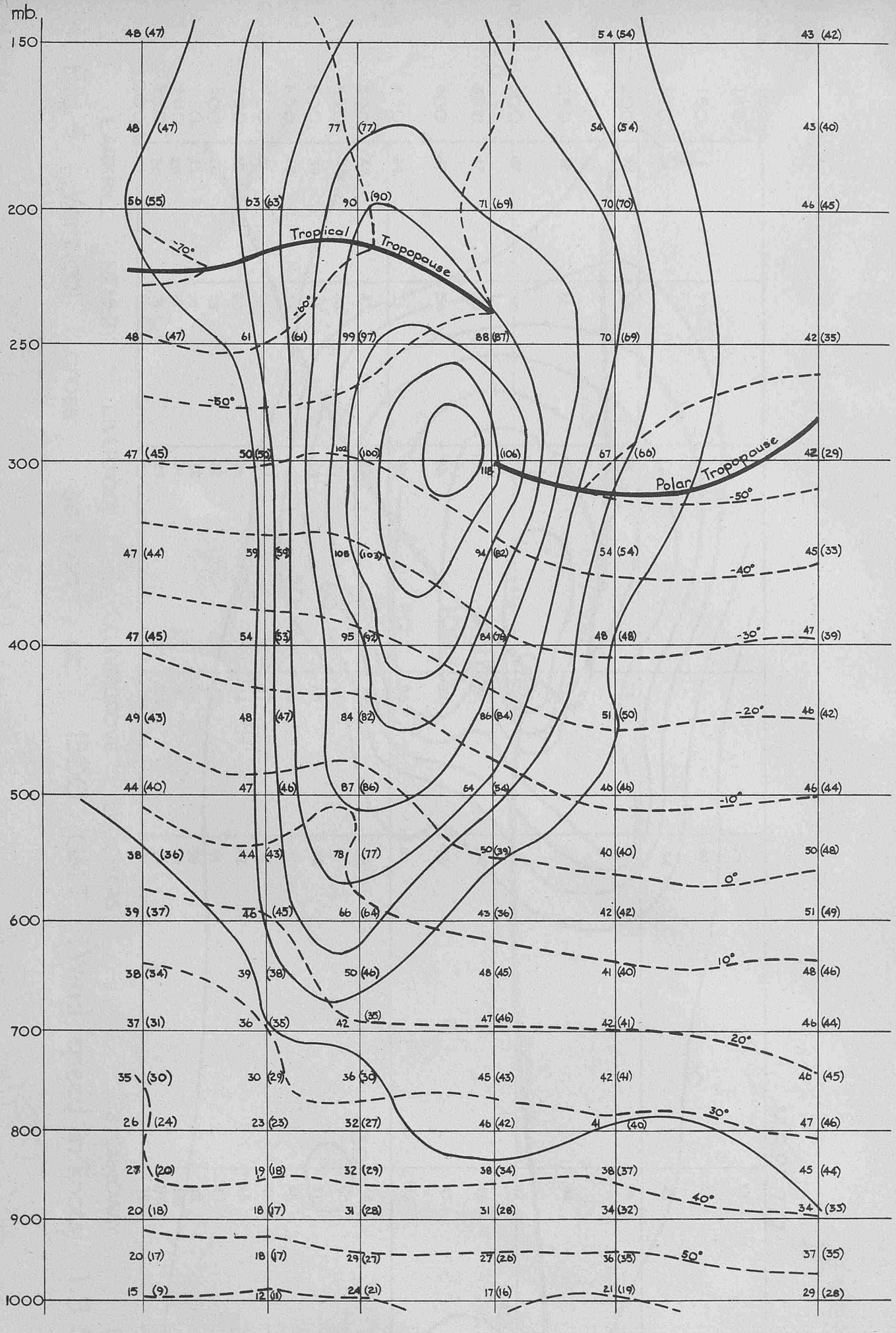

Figure 3. Vertical cross section Larkhill - Stornoway at 1500 G.M.T.

Explanation: Isotherms are shown by broken lines with temperatures in °F. entered against them. The speed of the actual wind is shown by the figures at each level above the stations. The figures in brackets show the component of wind at right angles to the plane of the paper.

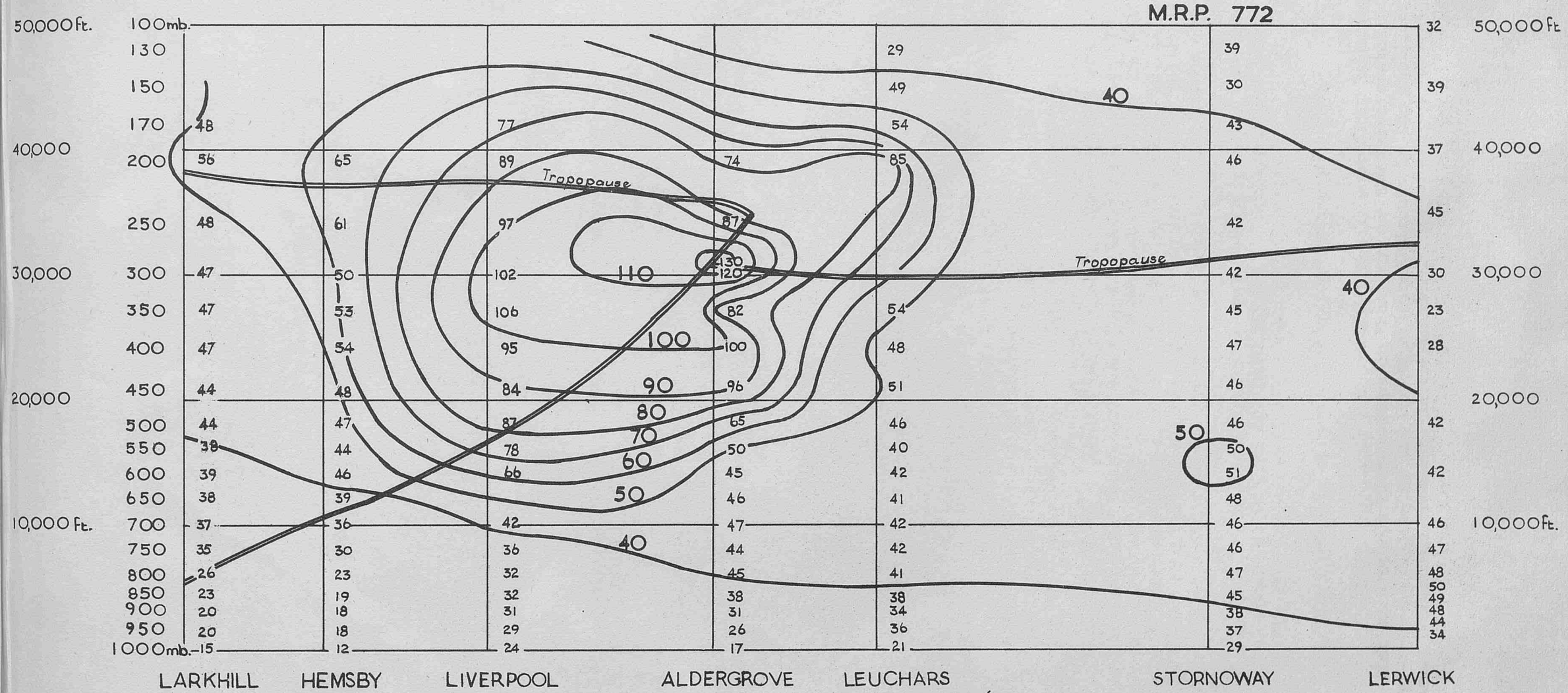

Fig. 4 Vertical cross section at 1500 G.M.T (Wind speed in knots) 1.9.52

M.R.P. 774
S.C.I./72
14th November, 1952.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No..... by Mr.....
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An improved automatic frost-point hygrometer.

by H. P. Palmer.

1. Introduction.

The ex-Elliott automatic hygrometer gave encouraging results during flight trials in December, 1951, as reported in M.R.P. 713. These results showed that an automatic hygrometer could give accurate measurements of humidity, and a speed of response to changing humidities which was considerably faster than could be obtained with a manually operated visual hygrometer. This encouraging performance was only obtained by making the hygrometer relatively complicated, (see block diagram in figure 1) and the instrument was correspondingly heavy and bulky. It could only operate from stabilised power supplies and it was very easily disturbed by electrical 'noise' in those supplies.

2. The ex-Elliott automatic hygrometer.

(a) Description.

In the ex-Elliott hygrometer a deposit of ice is accumulated on the top surface of a thimble of aluminium, and light scattered by this deposit is detected by a photomultiplier. The thimble is continuously cooled by conduction, and it is heated by passing a current through a resistance wound on the side of the thimble (figure 2). Variation of this current changes the thimble temperature, and the current is controlled by the photomultipliers, through electronic and magnetic amplifiers, so as to keep a constant deposit of ice on the thimble. The thimble temperature is measured by a thermocouple, whose output is amplified and indicated upon a milliameter calibrated in degrees centigrade.

When the thimble is ventilated with an airstream of constant water content the hygrometer comes to equilibrium with the thimble at a steady temperature, the frost-point, maintaining a constant deposit of ice. If the hygrometer reverts to these equilibrium conditions after some disturbance, it is said to be stable. With this design of thimble shown in figure 2, the stability is impaired by a delay of the order of one second between the occurrence of a change in the heater current, and the resultant change in the temperature of the thimble surface. This delay makes the hygrometer too slow to reach stability when the deposit is of dew, and stability with deposits of ice is only obtained with difficulty, and a considerable cost in complexity.

(b) The stability of automatic hygrometers.

An automatic hygrometer has nothing directly equivalent to mechanical friction to damp out fluctuations or "hunting" of the servo system. It can however be stabilised by the use of a "rate" circuit in the amplifier or by thermal conduction processes in the thimble.

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(c) Programme of work.

The results of the flight test (loc.cit. M.R.P.) demonstrated the advantages of an automatic hygrometer of rapid response but it was felt that the possibilities of the design had been exhausted. It was decided to investigate methods of reducing the thermal delay in the thimble, and if possible to obtain stability from the thermal effects, without using a rate circuit.

3. The use of a germanium thimble.

(a) Introduction.

In an attempt to reduce the thermal lag in the hygrometer thimble the author has investigated several methods of developing heat near the thimble surface. (The maximum power available at the thimble of the ex-Elliott hygrometer is about 10 watts). The methods which did not give satisfactory results are described in appendix 1. These experiments suggested that some of the methods would be more successful if the thimble were constructed of a material which had a good thermal conductivity but a high electrical resistivity so that the heat could be developed by passing the heating current directly through the thimble. Unfortunately such a combination of properties is not found among metals. After a discussion of these experiments, Mr. E.L. Simmons of the Clarendon Laboratory suggested that a semi-conductor such as pure silicon possessed both these properties. It was learned subsequently that high purity germanium is more readily available. Through the kindness of Mr. T.C. Kinman a thimble of high purity germanium was produced at the research laboratories of B.T.H. Rugby, and was presented to us.

(b) The properties of high purity germanium.

Pure germanium is a very hard, very brittle, grey substance which takes a high polish. Its thermal conductivity is 0.2 cals./cm. sec. deg. C and its electrical resistivity is 40ohms/cm. These are to be compared with the corresponding values of 0.5 and 2.8×10^{-6} for pure aluminium. The first germanium thimble had an electrical resistivity of 42 ohms/cm. at room temperature; this makes heating by passing a current through the thimble quite easy while for pure aluminium with a resistivity of 2.8×10^{-6} ohms/cm. it is quite impracticable. The resistivity of this sample of germanium increased rapidly as it was cooled, to about 200 ohms/cm. at -80°C. The resistivity of the sample is reduced if it is illuminated with light containing near infra-red radiation, to which it is transparent. Germanium may be nickel plated, and electrical and thermal contacts may be made by soft soldering to the areas of plate.

(c) The mounting of the germanium thimble.

The thimble was of the same dimensions as the aluminium thimble in the ex-Elliott hygrometer, as shown in figure 3. Two power contacts were plated on to the thimble as shown, of such a shape that the highest current density (lowest resistance path) would be across the top of the thimble between those parts of the plated contacts on the bevel of the thimble. The power leads, of tinned copper 1 amp. fuse wire, were soldered on to the small areas of each contact on the sides of the thimble. At room temperatures the resistance between the power leads was about 300 ohms when the thimble was dark, and about 150 ohms when the thimble was illuminated evenly with the light from the tungsten filament bulb in the hygrometer. When the thimble was at -80°C the corresponding figures were 1500 ohms and 800 ohms. The variation of thimble resistance with temperature results in some mis-match in the heater circuit design and reduces the heater power obtainable at low temperatures. This is not serious as the power required is also reduced at low temperatures. As regards the photo conduction effects, provided the disposition of the electric contacts and of the illuminated areas is correctly arranged the photo conduction increases the proportion of the heat which is developed close to the surface and this improves the stability.

/For

For the first experiments the thimble was soft soldered on to the end of an O B.A. brass rod. On two occasions the thimble was found to be fractured after experiments, probably because of stresses which were the result of the different coefficients of thermal expansion of germanium and brass. (Brass $\alpha = 19 \times 10^{-6}$; Ge, $\alpha \approx 9 \times 10^{-6}/°C$). In subsequent experiments thermal stresses were avoided by mounting the thimble on the end of a bunch of copper wires (figure 4) which gave a conduction path equivalent to the brass rod.

4. The performance of the hygrometer with a germanium thimble.

(a) Stability with a rate circuit.

The first experiments with the germanium thimble showed that the stability and speed of response of the hygrometer were both much improved. The thermal integrating effects were found to be sufficiently powerful to stabilise the hygrometer effectively, even when the rate circuit was disconnected. This has made possible a great simplification of the hygrometer amplifier and power supplies. The ex-Elliott hygrometer required a two-stage D.C. amplifier, with stabilised H.T. supply and an independent supply for the cathode heaters. There was also the magnetic amplifier with an independent transformer. Because the loss of D.C. gain in the rate circuit was avoided in the hygrometer with a germanium thimble only a single-stage D.C. amplifier was necessary, and this and the magnetic amplifier could be operated satisfactorily from a single, unstabilised, power supply.

(b) Operation on ice deposits.

The improved speed of response may be illustrated by comparative figures for the ex-Elliott hygrometer with an aluminium thimble, and the hygrometer with a germanium thimble. In each case the thimble was ventilated with an airstream of frost-point -50°C, and the thimble was initially clear of deposit and at 0°C.

	Aluminium thimble.	Germanium thimble.
Thimble cools to -50°C.	20 secs.	20 secs.
Hygrometer stabilises.	20 secs.	8 secs.
'Search and find' (Total)	40 secs.	28 secs.

(c) A flight trial.

The hygrometer with a germanium thimble was flown in a Hastings aircraft of M.R.F. on July 29th, 1952. It operated very satisfactorily for 30 minutes at about 15000', but the frost-point was steady, so the rapid response was not tested. On the descent the thermocouple amplifier developed an intermittent fault, and, as reported below, the hygrometer went out of action as soon as a dew deposit was formed.

(d) Operation with deposits of dew.

The thermal lag in the germanium thimble is sufficiently small for the hygrometer to be reasonably stable when working with deposits of dew. The ex-Elliott hygrometer was unstable, with the dew deposit alternately growing and evaporating completely, and with the thimble temperature oscillating through 5-8°C. Unfortunately the dew deposits on the germanium thimble were rapidly electrolysed because of the relatively large voltages which appeared across the surface of the thimble. (About 25 volts A.C. from side to side.) When electrolysis occurred white germanium salts were formed, the thimble surface was etched, and the hygrometer quickly went out of action.

It was found that a thin layer of silicone grease applied to the surface was sufficient to insulate the dew deposit when the instrument was used in the laboratory. The hygrometer has operated on dew continuously for an hour with such insulation. For the flight trial the surface was similarly insulated,

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but electrolysis occurred as soon as a dew deposit was formed, and the hygrometer went out of action.

It seems probable that electrolysis can be prevented by applying a coat of a suitable varnish to the thimble surface, and by using germanium of rather lower resistivity, so that the voltages across the thimble are reduced. For further experiments a germanium thimble of resistivity 15 ohms/cm. has been supplied by B.T.H., again through the kindness of Mr. T.C. Kinman.

5. The use of resistance thermometry.

When the hygrometer thimble was heated by a resistance wound on the side of the thimble it was not practicable to find space to wind a resistance thermometer on the thimble as well. Now that a heater resistance is no longer required it should be possible to use a resistance thermometer. This would give a great gain in thermometric power over the thermocouples used at present, and it would then be possible to dispense with the complication of a thermocouple amplifier.

In the laboratory experiments the two thermocouples which were soldered to the germanium (figure 3) were used to measure the temperature gradient between the top and bottom of the thimble when the hygrometer was operating. The gradient was about 6°C when the thimble was at room temperature and about 4°C when the thimble was at -60°C. This gradient should depend only on the temperature of the thimble, so that it should be possible to calibrate a resistance thermometer to indicate the temperature of the thimble surface.

6. Conclusion.

The improved automatic frost-point hygrometer described in this paper is very much simpler than the ex-Elliott hygrometer (compare figures 1 and 4) and has a more rapid response. It should now be possible to develop an instrument which is correspondingly lighter, and in which the thimble temperature is measured by a resistance thermometer. Such an instrument could probably be flown from a balloon, and so make possible the measurement of stratospheric humidities to much greater heights.

The rapid response of the instrument, with the reduced thermal lag at the thimble, enables the hygrometer to operate stably with deposits of dew. Provided that electrolysis of the deposit can be prevented completely, it should now be possible to construct a recording hygrometer for use in cloud investigations below the freezing level.

7. Acknowledgements.

We are very grateful to Mr. E.L. Simmons of the Clarendon Laboratory for pointing out the advantages of using a semiconductor material for the thimble, and to Mr. T.C. Kinman of B.T.H. for supplying the high purity germanium thimbles.

Appendix. Methods of reducing the thermal lag in the thimble.

Several methods of developing power near the thimble surface were considered, and trial apparatus was built to try three of these. In the following notes the thermal delay in the aluminium thimble with resistance heater is denoted by L.

1. Radio heating.

This method has been used in American work by Suomi and Barrett and has some advantages. In the American work the inductor coil is mounted concentric with the cylindrical thimble, and the power is developed in a thin skin on the sides of the thimble. The power may be developed in a thicker skin if the thimble has a higher resistivity or if the radio frequency is reduced. In this latter case the inductor coil must carry heavy currents, and is correspondingly bulky. (With a frequency of 20 kc, and an aluminium thimble, the

/inductor

inductor current is estimated to be of the order 100 amps). It seems impossible to develop the power near the top surface of the thimble without interfering with the optical arrangements for detecting the deposit. Radio heating might reduce the lag to perhaps $\frac{1}{2}$L, but it was felt that this method was not sufficiently attractive to justify experimental work.

2. Evaporated films.

Consideration was then given to methods of forming the thimble surface of a thin electrically insulated metallic film of high electrical resistivity. It was hoped to develop the power by passing a current through the film. Films were formed by vacuum evaporating metals on top of the anodised coat of aluminium thimbles. It was found that the films were not sufficiently adherent to the supporting anodising to stand the cleaning and polishing which a hygrometer thimble must receive. The most promising results were obtained with Rhodium films, and improved techniques of evaporation might give usable films.

3. Dielectric heating.

The next method attempted was to form the thimble surface of a cap of aluminium foil glued to the anodised aluminium thimble. Power was developed by dielectric heating of the glue, using the foil and the thimble as the two plates of a 'lossy' condenser. This method depends on the glue having a large power factor, and although several dielectrics were tried, the highest power factor measured was 0.05. It was possible to develop about 2 watts in this thimble by applying 500 volts R.F. across the condenser, but these large voltages rendered the method unattractive unless a dielectric with a larger power factor could be found.

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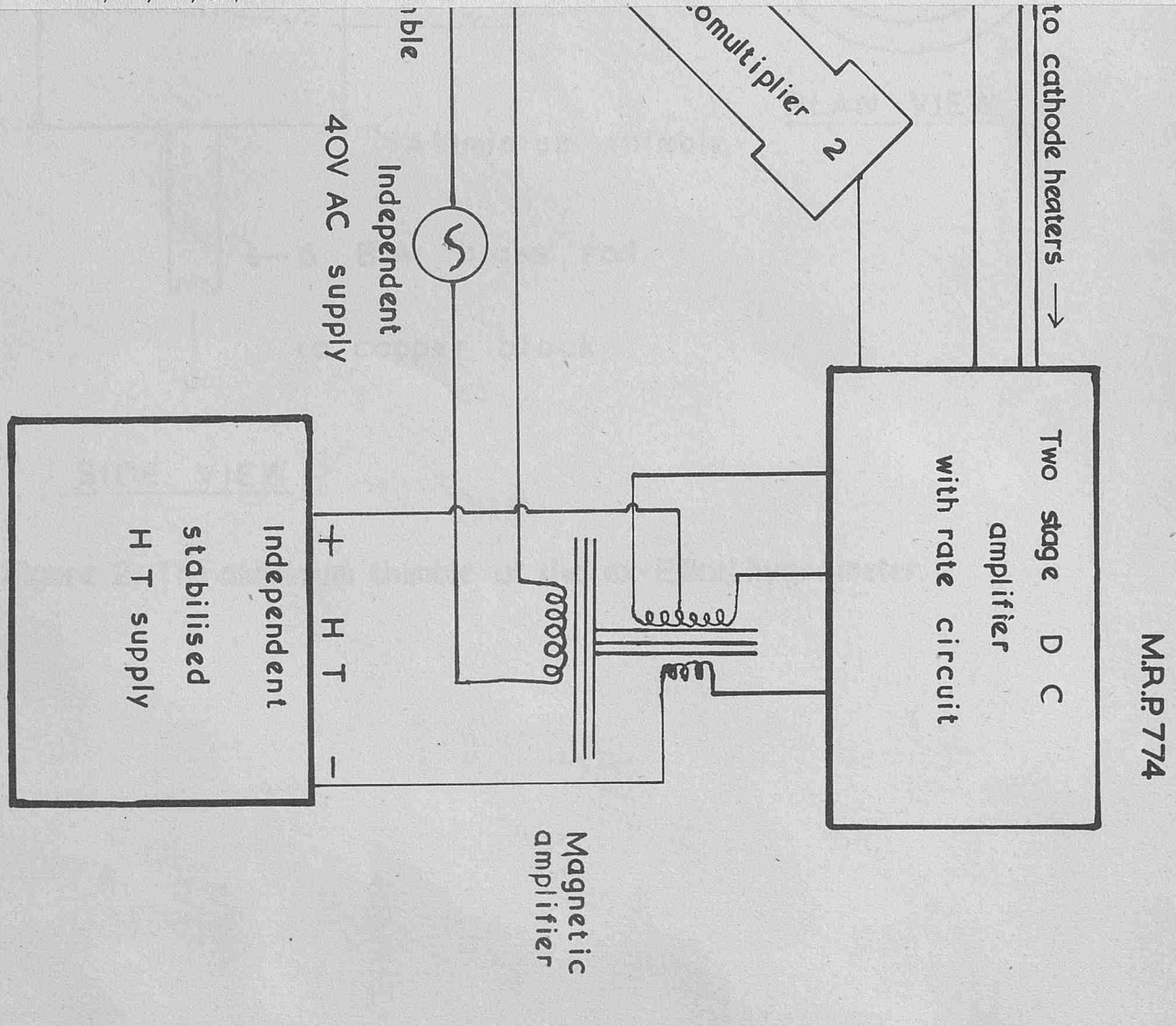

natic hygrometer as flight-tested.

4. Messrs. Elliott Bros. have developed an improved automatic hygrometer in which the thimble forms part of the single turn secondary winding of an iron cored transformer. With this system the thermal lag in the thimble is probably reduced to about $\frac{1}{2}$L. Some experiments were carried out in an attempt to reduce the lag still further by this method. An experimental assembly was made and with such a thimble it was possible to develop a power of about 2 watts in the thimble. There was serious leakage of magnetic flux round the sides of the thimble, and the power could not be increased further.

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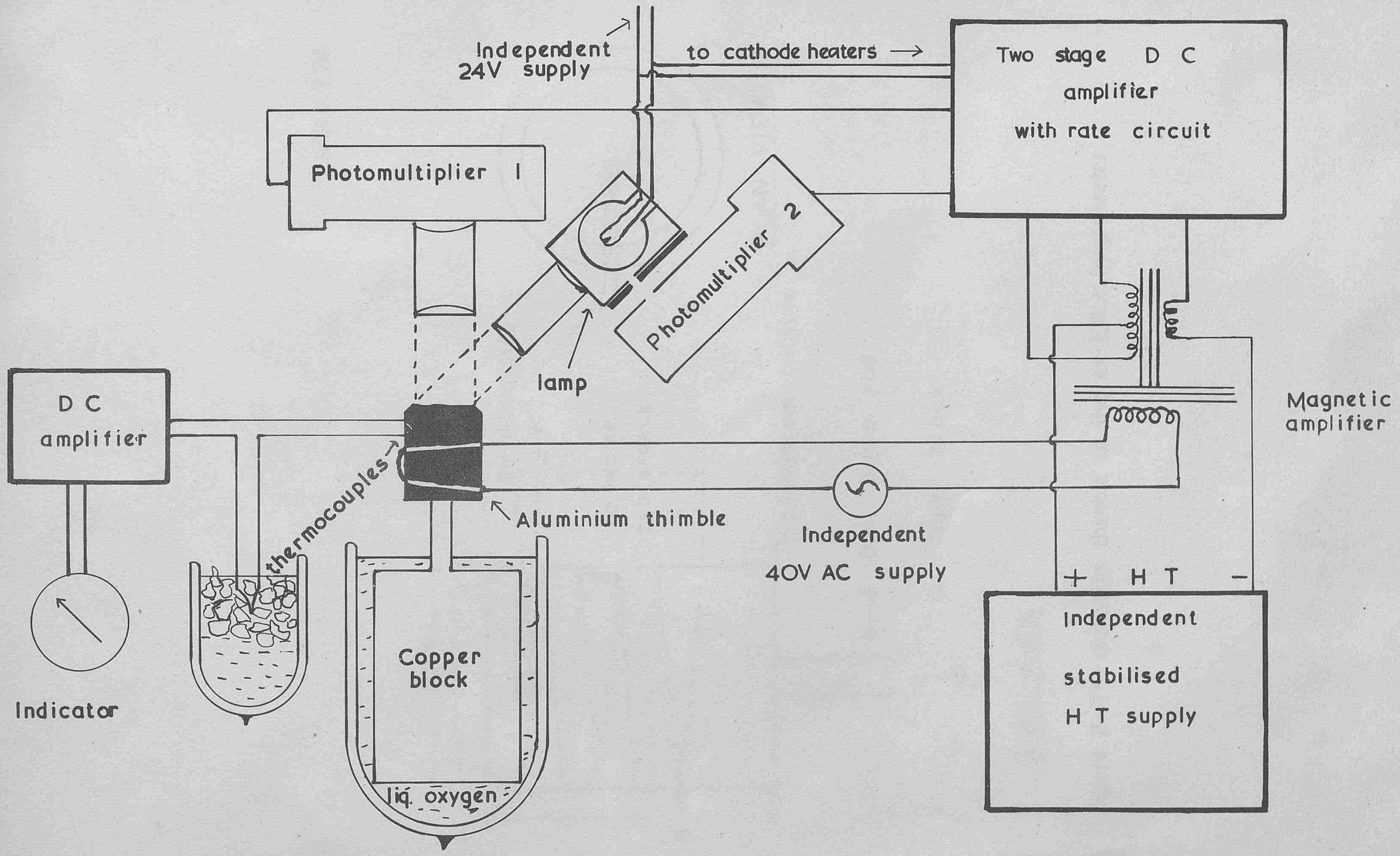

Figure 1. Schematic diagram of the ex-Elliott automatic hygrometer as flight-tested.

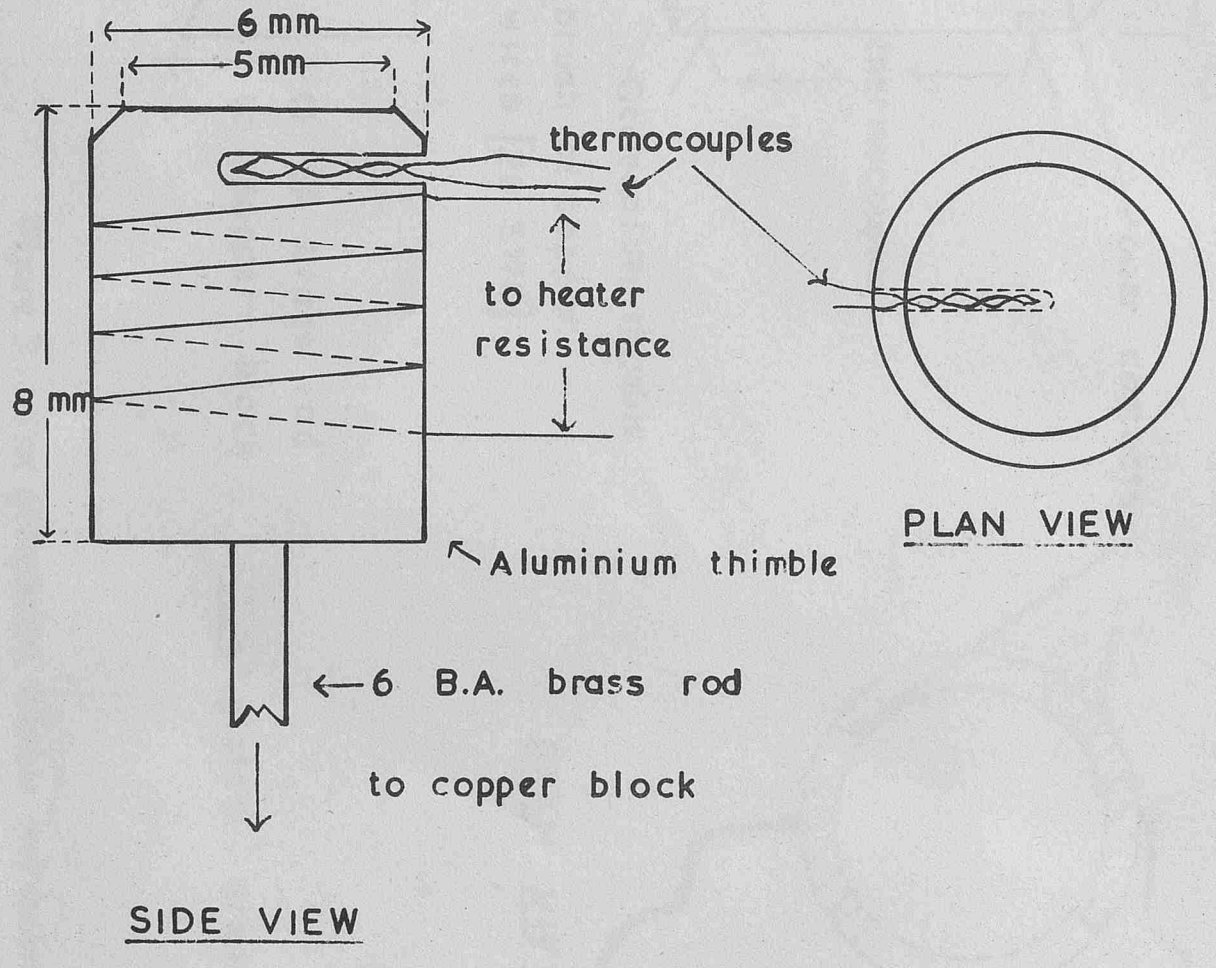

Figure 2. The aluminium thimble of the ex-Elliott hygrometer.

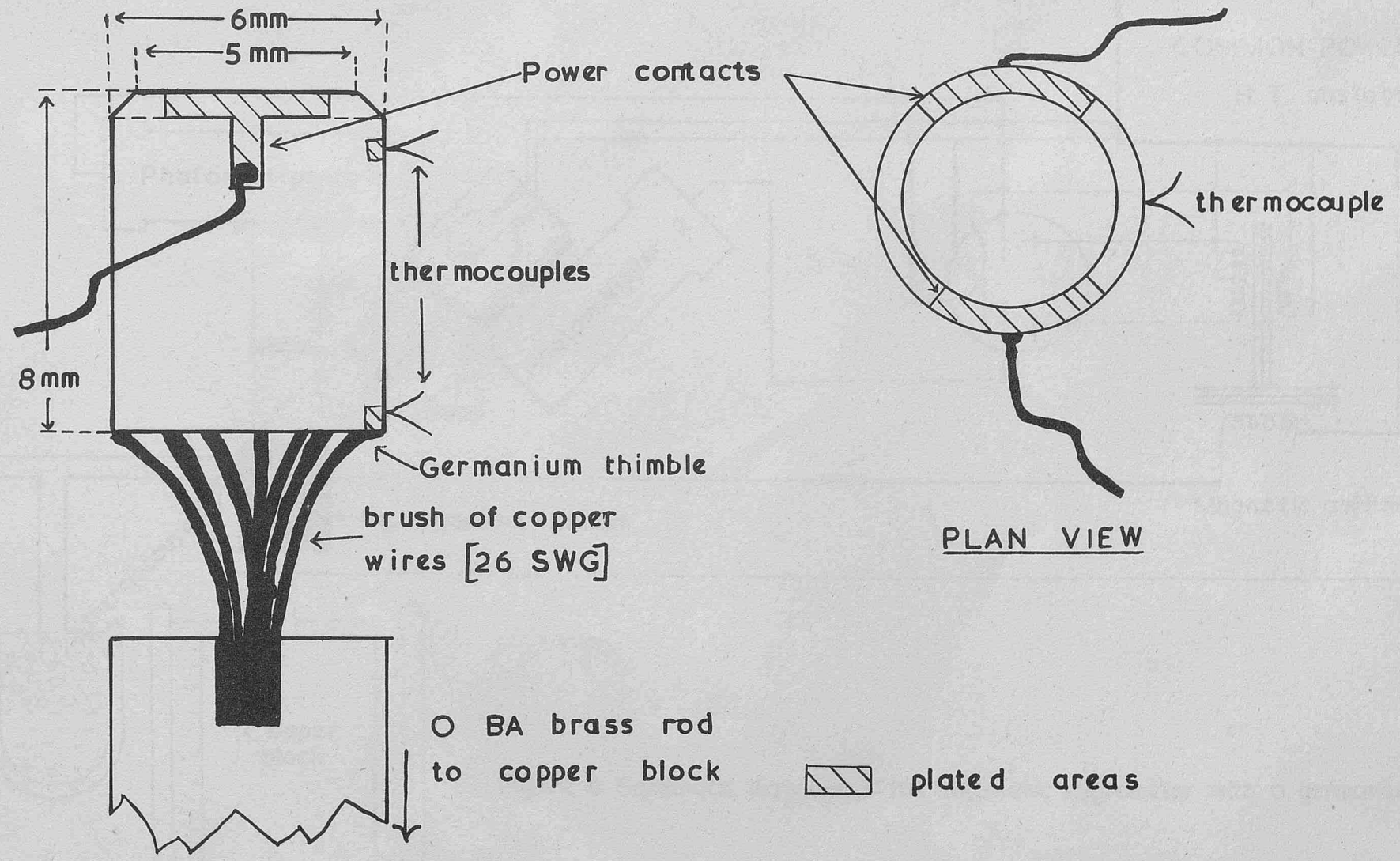

Figure 3. The germanium thimble and mount.

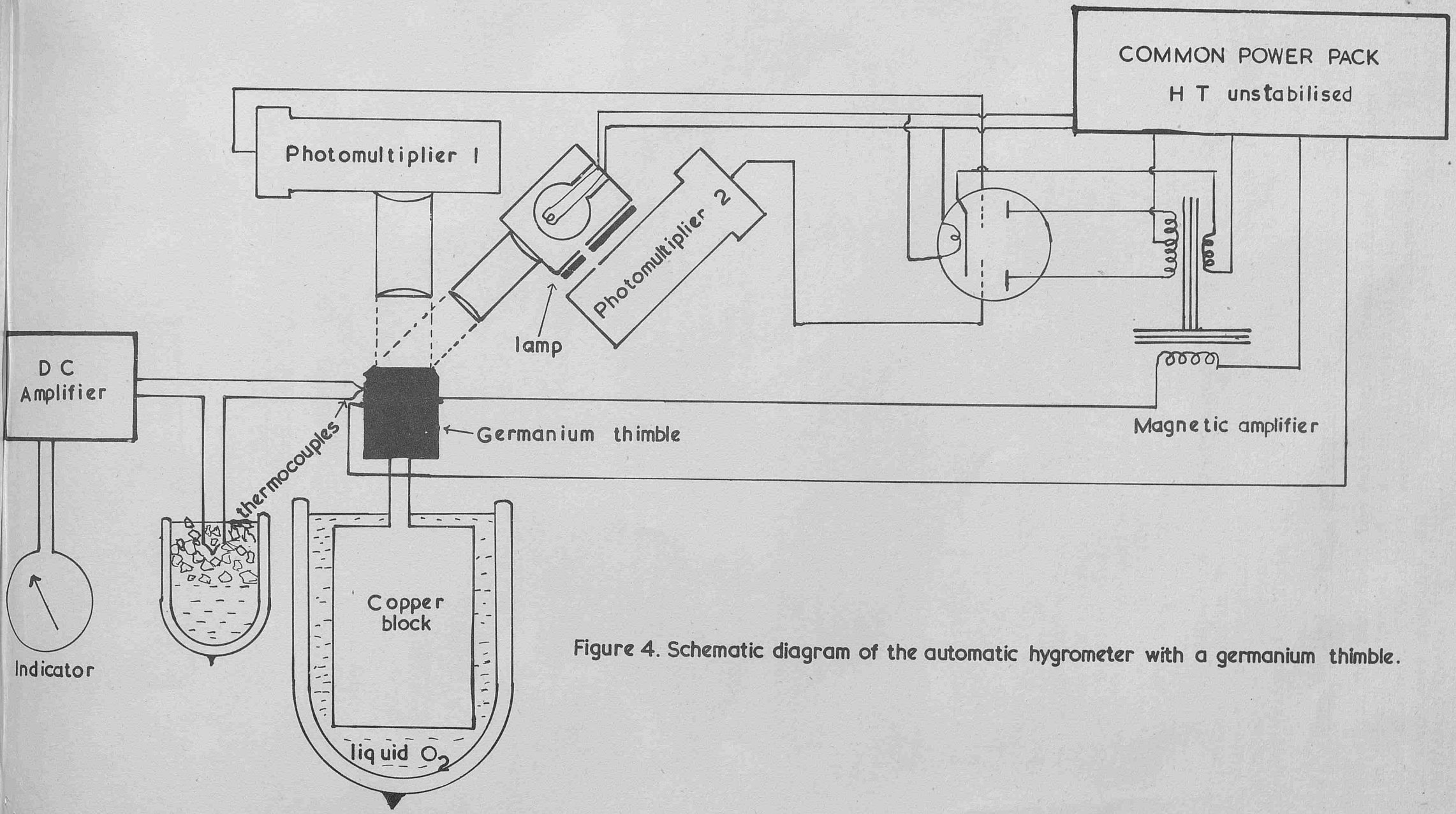

Figure 4. Schematic diagram of the automatic hygrometer with a germanium thimble.

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M.R.P. 775
S.C.II/132
18th November, 1952.

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AIR MINISTRY
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The effect of atmospheric inhomogeneity on the interpretation of vertical temperature soundings

by
J.S. Sawyer (Meteorological Office)

1. Introduction

The atmosphere is subject to oscillations and fluctuations ranging in period from a fraction of a second to several days and in scale from a few cm. to 1000 Km. or more. For the purpose of synoptic weather analysis and forecasting attention is concentrated on weather systems with a minimum horizontal extent of 200-300 Km.; smaller scale irregularities affect only one reporting station at a time and therefore cannot be incorporated in the analysis.

Frith (1950) has demonstrated the existence of temperature fluctuations of up to 2°F. over horizontal distances of 50 Km. These disturbances are too small to be incorporated in synoptic weather analysis, but nevertheless the temperature differences which arise in them are sufficient to be of interest to the synoptic analyst. The departures of the temperature from the true mean value for an area of the order of 100 Km. square, must be regarded by the analyst as an error due to the unrepresentative nature of his sounding.

Relatively small-scale temperature fluctuations become of particular significance in the analysis of vertical soundings of temperature. The small-scale temperature variations may introduce changes in lapse rate which are quite unrepresentative of the area around the sounding station, but which may simulate frontal lapse rate discontinuities or lead to false deductions regarding air mass stability. It is the purpose of the present note to investigate the limitations imposed on the interpretation of vertical soundings by the existence of such temperature fluctuations.

A direct method of carrying out such an investigation would be to compare individual temperature soundings with the mean temperature sounding representative of an area of radius 50 Km. or so around the sounding station. Unfortunately there is no method of obtaining mean temperatures for a horizontal area of 100 Km. extent in the free atmosphere simultaneously at a large number of levels without the use of a prohibitively large number of radio sondes or aircraft. We shall therefore approach the problem from a different point of view and in the following paragraphs an estimate is made of the effect of the small-scale temperature fluctuations in modifying some tephigrams based on simple air-mass and frontal conditions. An examination of these reveals the magnitude of the spurious irregularities in lapse rate.

For our purpose information is required on the magnitude of the temperature fluctuations and on the vertical scale of the disturbances. This has been taken from the original records of the Meteorological Research Flight which carried out a large number of flights during 1948 and 1949 on the lines of those described by Frith (1950).

2. The standard deviation of temperature

From the records of the Meteorological Research Flight, 34 cases were found in which a horizontal grid had been flown within an air mass (i.e. not within a frontal zone). The flights were conducted on the lines described in M.R.P.583. In some cases more than one grid was flown on a particular occasion usually at different levels.

/From

From each grid flight the standard deviation of temperature from the mean for the flight was calculated. Three of the flights were conducted in an inversion or markedly stable layer. These gave standard deviations of temperature of 0.79, 0.77 and 0.78°F., values larger than obtained on the remaining flights. The average standard deviation for the remaining flights was 0.51°F. and the standard deviation of the individual s.d's. about this mean was 0.16°F.

A reasonably representative value of the standard deviation of temperature over an area of some 100 Km. therefore seems to be about 0.5°F. in normal air masses and about 0.75°F. in inversion or other very stable layers.

3. The vertical correlation of the temperature fluctuations

In order to discuss the effect of small-scale temperature fluctuations on the appearance of a temperature sounding curve, it is necessary to know, not only the magnitude of the temperature fluctuations, but also the vertical scale of the temperature variations. This is best examined by means of the correlation co-efficients between the temperatures at pairs of levels with various vertical separations.

Flights have been made by Meteorological Research Flight aircraft to examine the vertical scale of the temperature fluctuations. In some of these two aircraft were used flying simultaneously along parallel tracks with vertical separation from zero to 2,000 ft. On other occasions one aircraft only was employed and this returned along a track as near as possible vertically above the air through which it flew on the outward flight. Data are available from some 14 pairs of levels, and the correlation co-efficients between the temperatures at each pair of levels have been calculated. The correlation co-efficients are plotted against the vertical separation between the levels in Fig.1(a).

As would be expected the correlation co-efficient decreases with increasing separation between the levels and becomes zero when the separation reaches about 2,000 ft. When two aircraft are employed flying as close together as possible the correlation co-efficient does not reach unity. This may be attributed to errors of observation and to temperature variations on a still smaller scale. If we assume that this difference from unity is due to random errors of observation, we may correct the remaining correlation co-efficients, and this has been done in Fig.1(b) to give a better representation of the true correlation between the temperature fluctuations at two levels.

Although, as would be expected, there is a good deal of scatter in Fig.1(b), nevertheless it seems that the variation of the correlation co-efficient with separation between the levels can be adequately represented by the curve superimposed on the diagram, and such a variation has been assumed in assessing the effect of the temperature fluctuations on vertical soundings.

4. The simulated temperature fluctuations

During the flight of a radio-sonde, observations of temperature, pressure and humidity are made in sequence so that a set of discrete observations of temperature are obtained referring to a set of levels separated by about 400 ft. The observed temperatures will differ from the true mean temperature representative of the area by an amount ε due to the small scale temperature fluctuations. The values of ε at consecutive observations will be correlated, and the correlation will decrease with increasing separation and become zero between observations separated by 2,000 ft. (i.e. five observations apart).

The next step in the investigation was to construct sets of numbers which could represent possible sets of values of ε (i.e. sets of departures of observed temperature from representative mean temperature at a sequence of levels 400 ft. apart). The conditions which had to be satisfied were:-

(a) The numbers should have a Gaussian distribution.
(b) Their standard deviation should be appropriate to the small scale temperature fluctuations.
(c) The autocorrelation of the series should be appropriate to the curve in Fig.1(b).

/Such

Such series have been constructed by first preparing random sets of numbers with a Gaussian distribution. The standard deviation has then been adjusted to be 0.5 or 0.75°F., appropriate to normal lapse or inversion condition respectively. From these sets a coherent series of numbers was prepared by forming running means of the form

$$\frac{1}{\sqrt{51}}\left\{2a_{n-2}+3a_{n-1}+5a_{n}+3a_{n+1}+2a_{n+2}\right\}$$

These coherent series should have autocorrelation co-efficients, $r_1 = +\cdot 82$, $r_2 = +\cdot 57$, $r_3 = +\cdot 24$, $r_4 = +\cdot 08$, $r_5 = 0$, appropriate to levels separated by 400, 800, 1200, 1600, and 2,000 ft. respectively.

In order to study the effect of such temperature fluctuations on the appearance of the tephigram, various idealised tephigrams were drawn appropriate to frontal and non-frontal situations. The temperature was read off at 400 ft. intervals and a set of values of ε taken from one of the coherent series and added term by term. The resulting temperatures were replotted on a tephigram and examined. The curves so obtained were more irregular than the usual temperature sounding used in upper air analysis. This difference is primarily a result of the smoothing introduced during the computation of the radiosonde results and their transmission to a central office. It was desirable therefore to simulate these processes also, before considering the effect of small-scale temperature fluctuations on aerological analysis.

It is unnecessary to describe the procedure of radiosonde computation in detail. It is sufficient to note that the radiosonde observations are first plotted against time and the curve is simplified by joining selected plots by straight lines, subject to the requirement that no intervening omitted point should differ by more than 0.5°F. from the straight lines drawn. The pressure and temperature are computed for the corner points and these are transferred to a temperature-log pressure diagram and again joined by straight lines. Interpolation on this diagram gives values for the levels which are reported to the central office and plotted on the final tephigram. These levels are at 50 mb. intervals, but additional significant points are reported wherever the error of reporting would otherwise exceed 0.5°F. This procedure has been closely simulated in the present experiment using the ideal lapse rates to which the simulated temperature fluctuations have been added. The tephigrams produced in this way will be referred to as the soundings " as reported", and the results will be described in the following paragraphs.

5. Simulated air mass soundings

Fig.2(a) shows four idealised tephigrams representing smooth lapse rates in homogeneous air masses. Fig.2(b) shows the result of modifying these soundings by the addition of a coherent series representing the effect of the small scale temperature fluctuations. (The standard deviations of the series used are indicated). Fig.2(c) shows the same soundings as they would have been reported.

It is interesting to note that the curves in Fig.2(c) are very similar in general form and in the nature of their irregularities to the curves actually plotted from radiosonde reports. A typical radiosonde plot is given for comparison in Fig.3.

In 3 out of the 4 soundings spurious stable layers have been introduced. The standard deviation of the series used in forming curves B and D (viz. 0.75°F.) is perhaps rather large for a homogeneous air mass but a stable layer also appears in curve A for which the standard deviation of the simulated temperature was only 0.5°F.

The irregularities of curves in Fig.2(c) are not large but they are sufficient to be misinterpreted by an analyst as minor air mass boundaries if he were to regard them as real, or to lead him to erroneous conclusions on the development of convective cloud.

6. Simulated frontal soundings

To illustrate the effect of small-scale temperature fluctuations on frontal soundings, an idealised frontal transition of moderate intensity has been studied. The same total air mass temperature change has been considered first as being linear over an interval of 100 mb. (curve I Fig.4(a)), second as being somewhat smoother over about the same pressure range (curve II) and thirdly as spread smoothly over a range of about 250 mb. Temperature fluctuations have been added to these idealised curves, and the soundings replotted as they would have been reported. The results are presented in Fig.4(b). (The standard deviation of the series used is indicated beneath the sounding).

The following points are of interest:-

(a) The sharp frontal transition of curve I has been little affected by the added fluctuation, but between 565 and 612 mb. another stable layer has been introduced which has an almost equally good claim to be treated as a frontal layer.

(b) The frontal transition is quite well defined in curve II "as reported". However, although in Fig.4(a) the boundaries of the frontal zone were deliberately smoothed off, in Fig.4(b) they appear to be quite sharp.

(c) The very smooth transition of curve III (Fig.4(a)) has been broken down into three distinct and rather sudden temperature changes. (Similar transitions are not uncommon on observed soundings and there has been some uncertainty as to whether the individual steps should be regarded as real frontal sub-zones).

7. Discussion of results

So far no mention has been made of the errors of the radio-sonde itself. Sheppard (1944) has estimated the probable error of reading as about 0.2°F. and the casual error of the temperature element itself as 0.6°F. (corresponding to standard deviations of 0.3 and 0.9°F.) A considerable part of the error of the element is probably systematic so far as a single sounding is concerned, and it seems probable that the effect of instrumental error on the sounding curves will be to add a random error with standard deviation about 0.4°F. to the individual readings. This will make the irregularity of the sounding curve somewhat greater than indicated in Figs. 2(c) and 4(b). It is evident that no useful purpose would be served by further reducing the purely random errors of the individual radiosonde readings because the irregularities due to real local temperature irregularities in the atmosphere limit the accuracy of the values which can be obtained.

In the preceding sections it has been demonstrated that minor irregularities in temperature soundings may easily arise from the known small scale temperature irregularities in the free atmosphere. These irregularities are large enough to be mis-interpreted as real phenomena of importance in frontal analysis or in restricting convection. It is difficult to lay down any specific rules for the reality of irregularities on a sounding curve, but it is clear that one must treat with caution any irregularity which does not depart by more than 1.5°F. from a smoothed curve following the sounding.

It is also apparent that it is quite impossible to decide whether a frontal zone is best represented by a linear transition with discontinuities of lapse rate at the boundaries, or by a gradually changing lapse rate; also that division of a broad frontal zone into two or more sub-zones is of doubtful validity unless the sub-zones themselves are very well marked on the sounding (departing by 2°F. or more from the smooth curve).

References

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P.A. Sheppard 1944 On the accuracy of the M.O. Radio-sonde temperature height and wind observations. Met. Office, London, S.D.T.M. 75, Part VII.

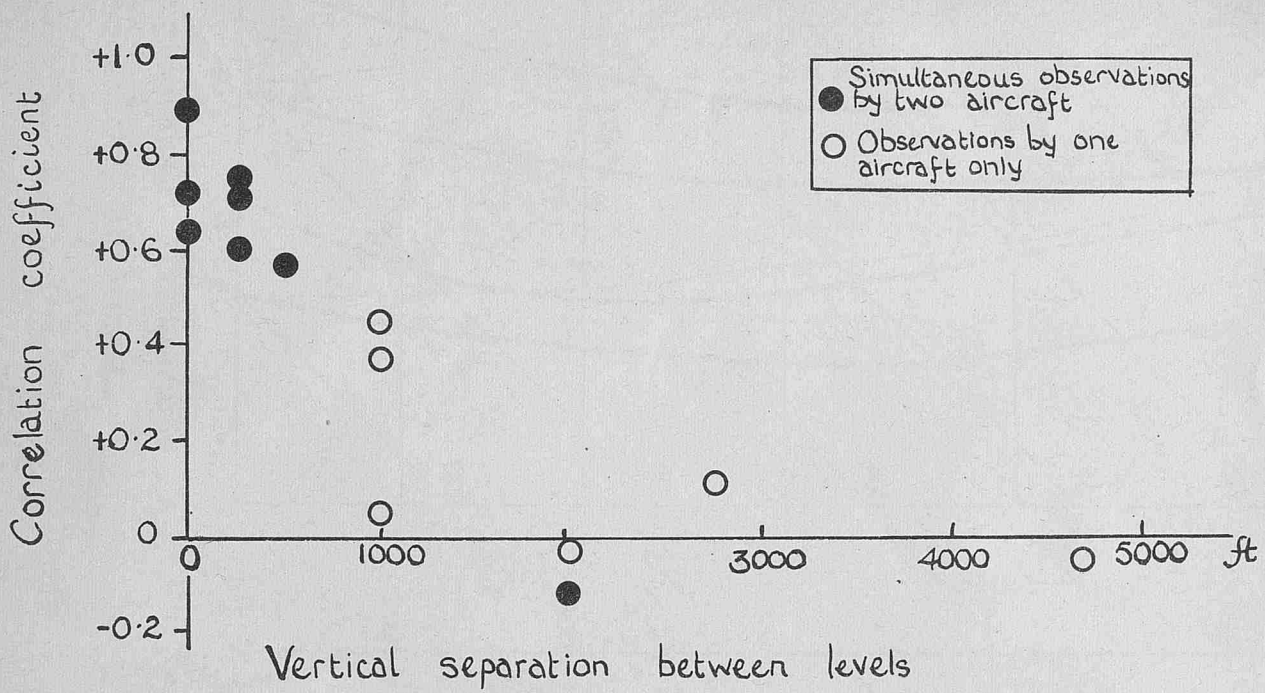


FIG I(a) Correlation coefficient between temperature at two levels over distances of 100 to 200 Km.

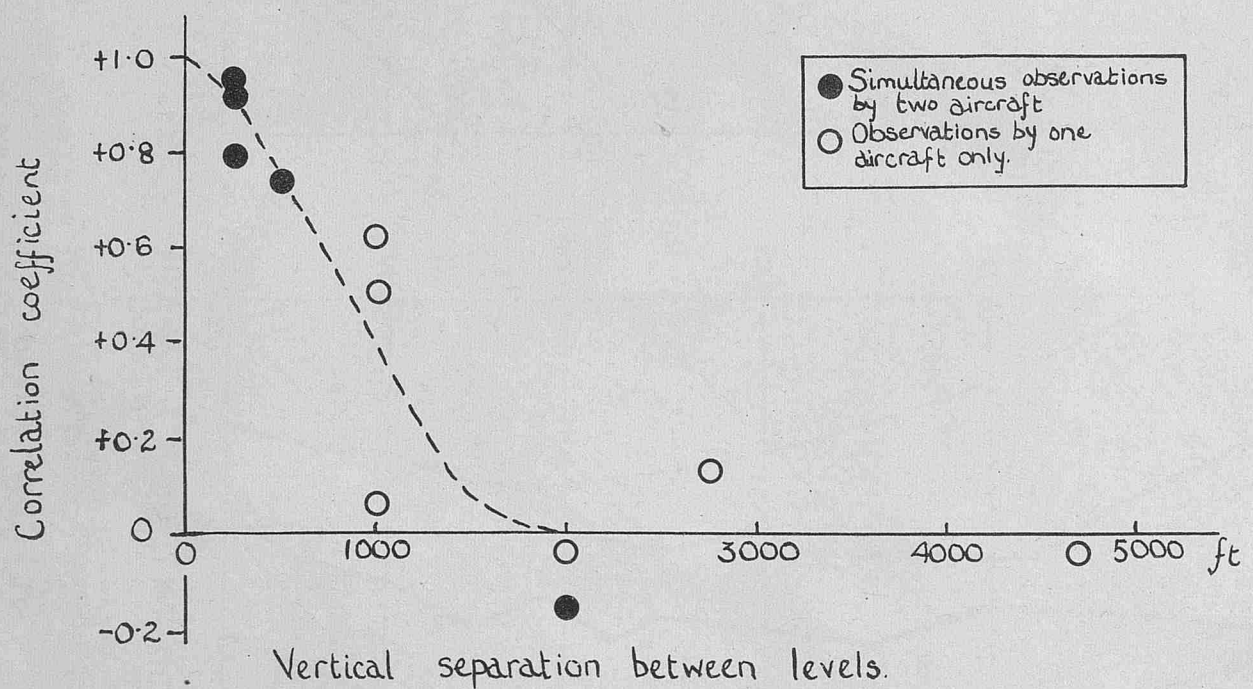


FIG I (b) Correlation coefficient between temperature at two levels (corrected for observational error.)

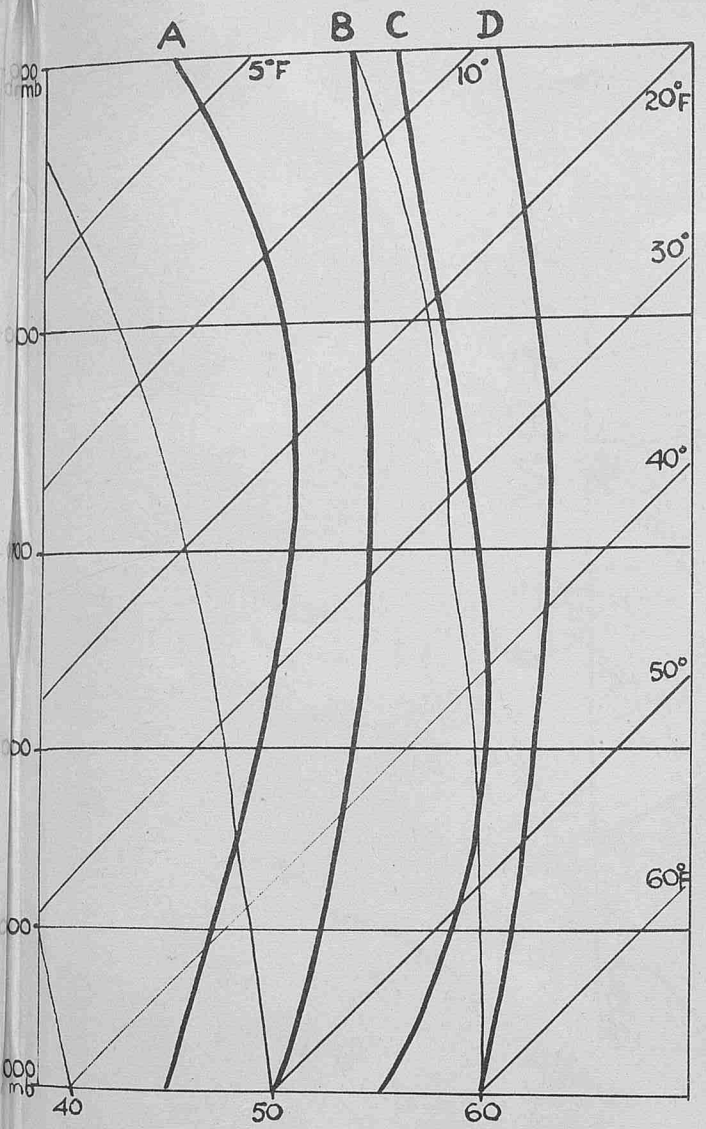
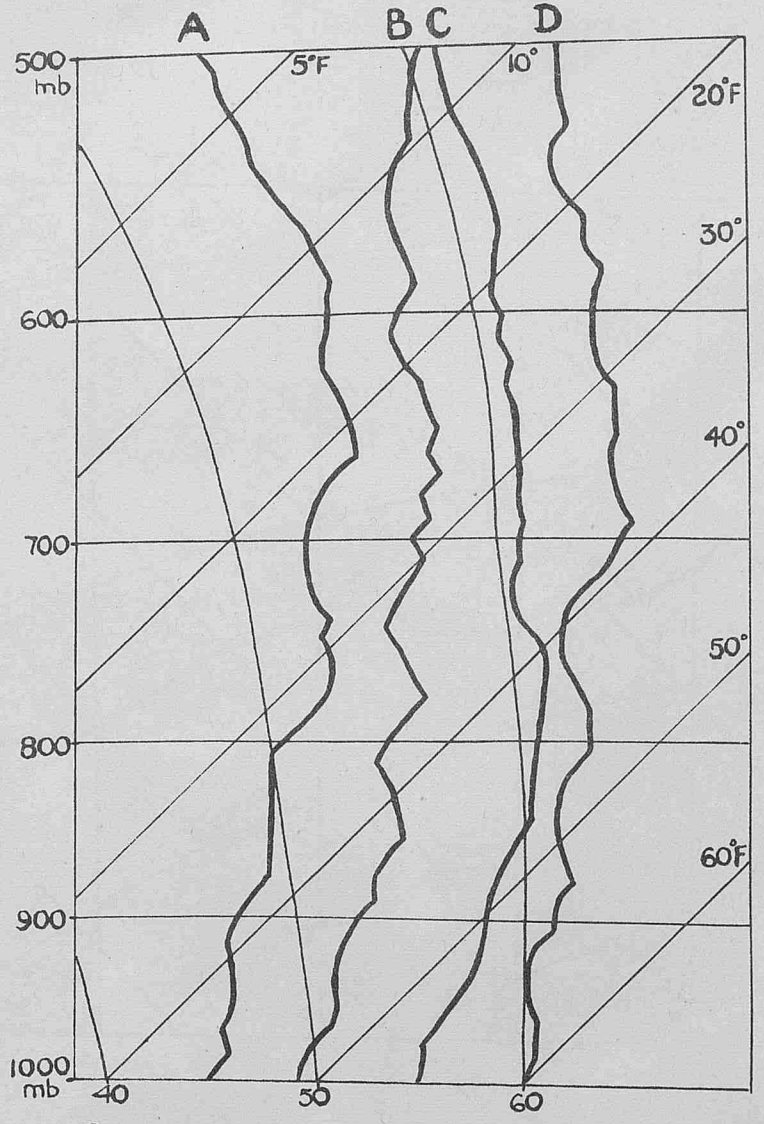
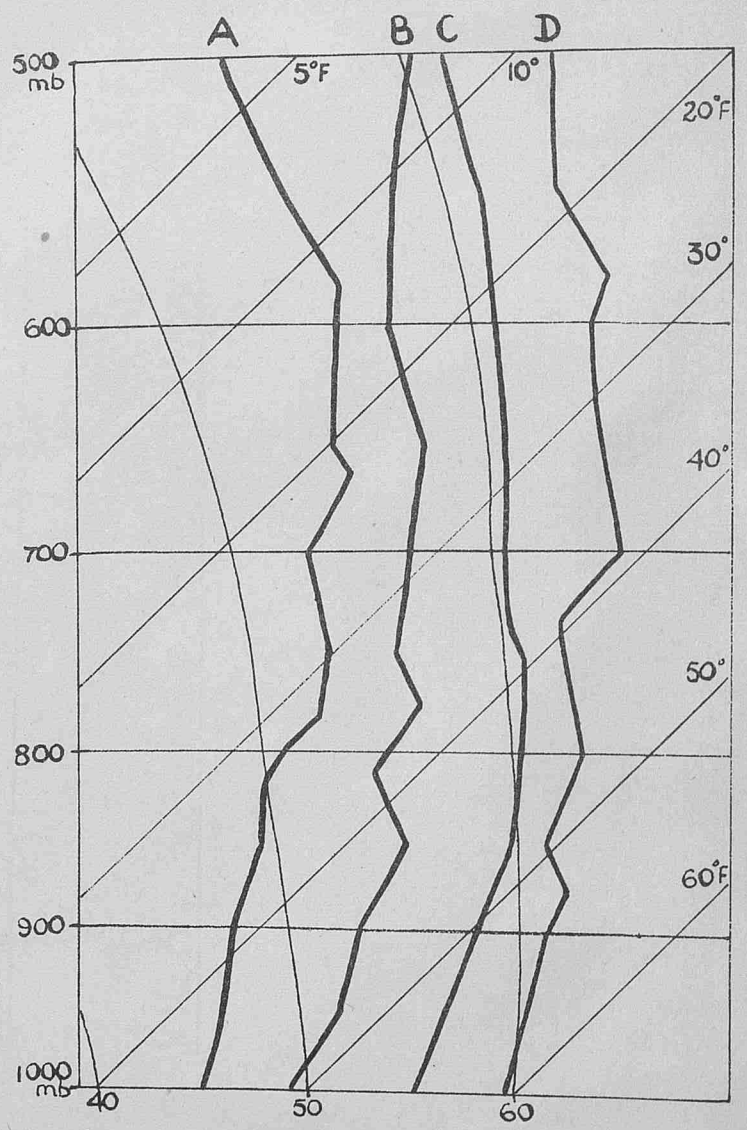


FIG 2 (a) - Idealised tephigrams for homogeneous air masses



Standard Deviation. 0·5°F 0·75°F 0·5°F 0·75°F

FIG 2 (b) - Idealised lapse rates modified by addition of coherent temperature fluctuations



Standard Deviation 0·5°F 0·75°F 0·5°F 0·75°F

FIG 2 (c) - Soundings of fig 2 (b) "as reported"

M.R.P. 775

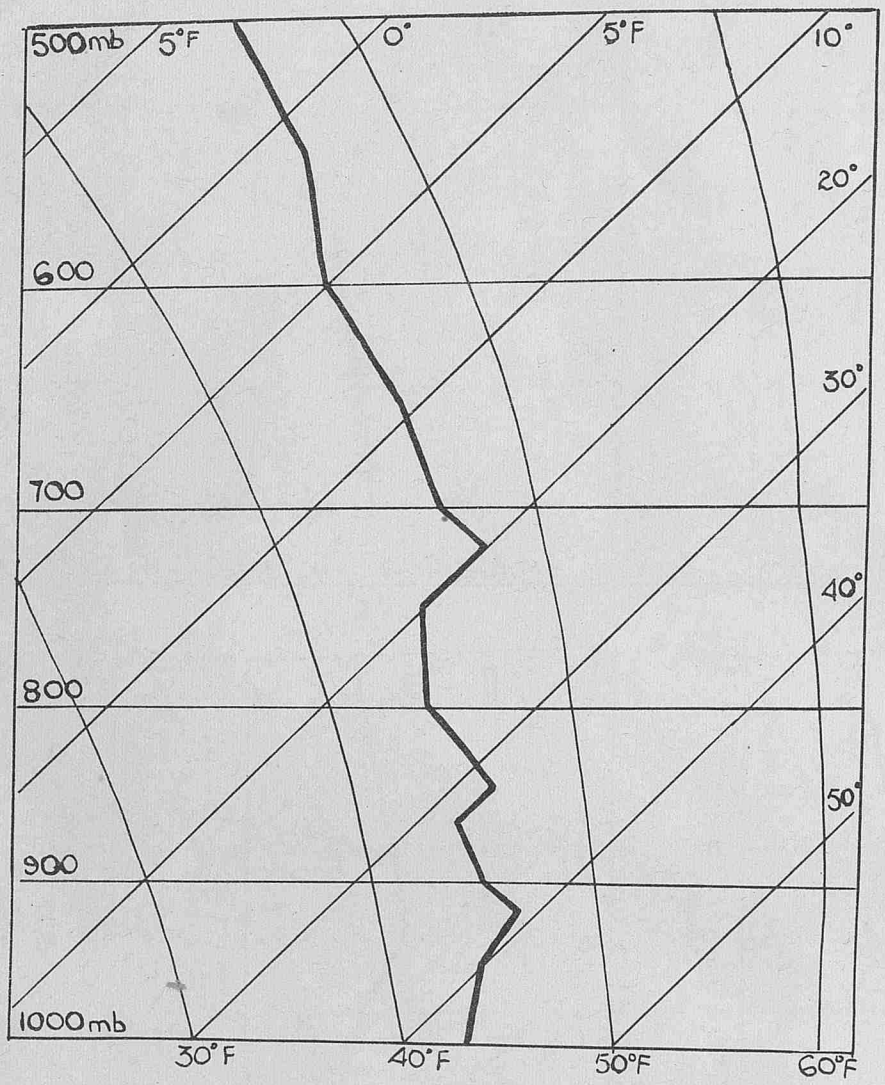


FIG 3 - Leuchars - 0300 9th Dec. 1950.

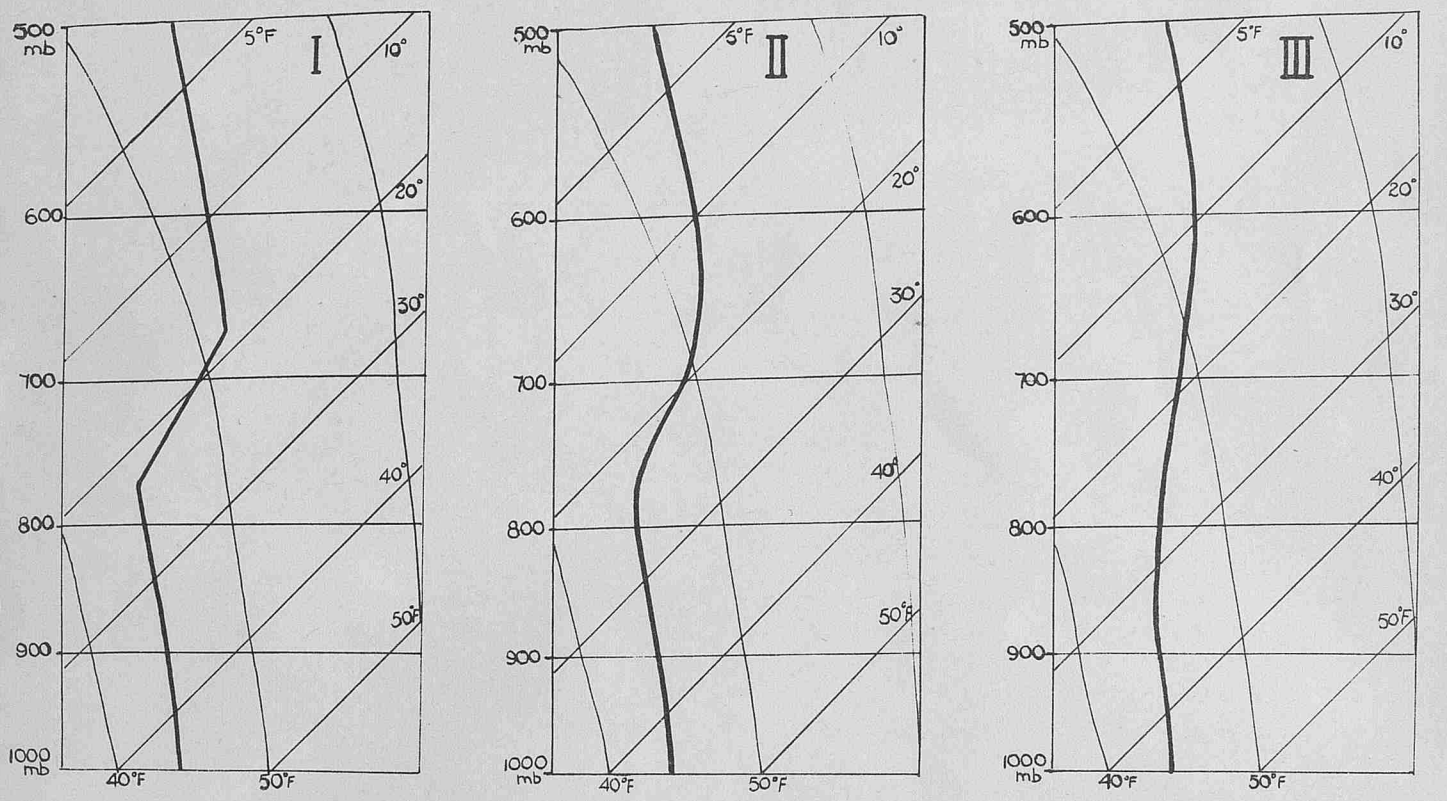


FIG 4(a) - Idealised soundings through fronts

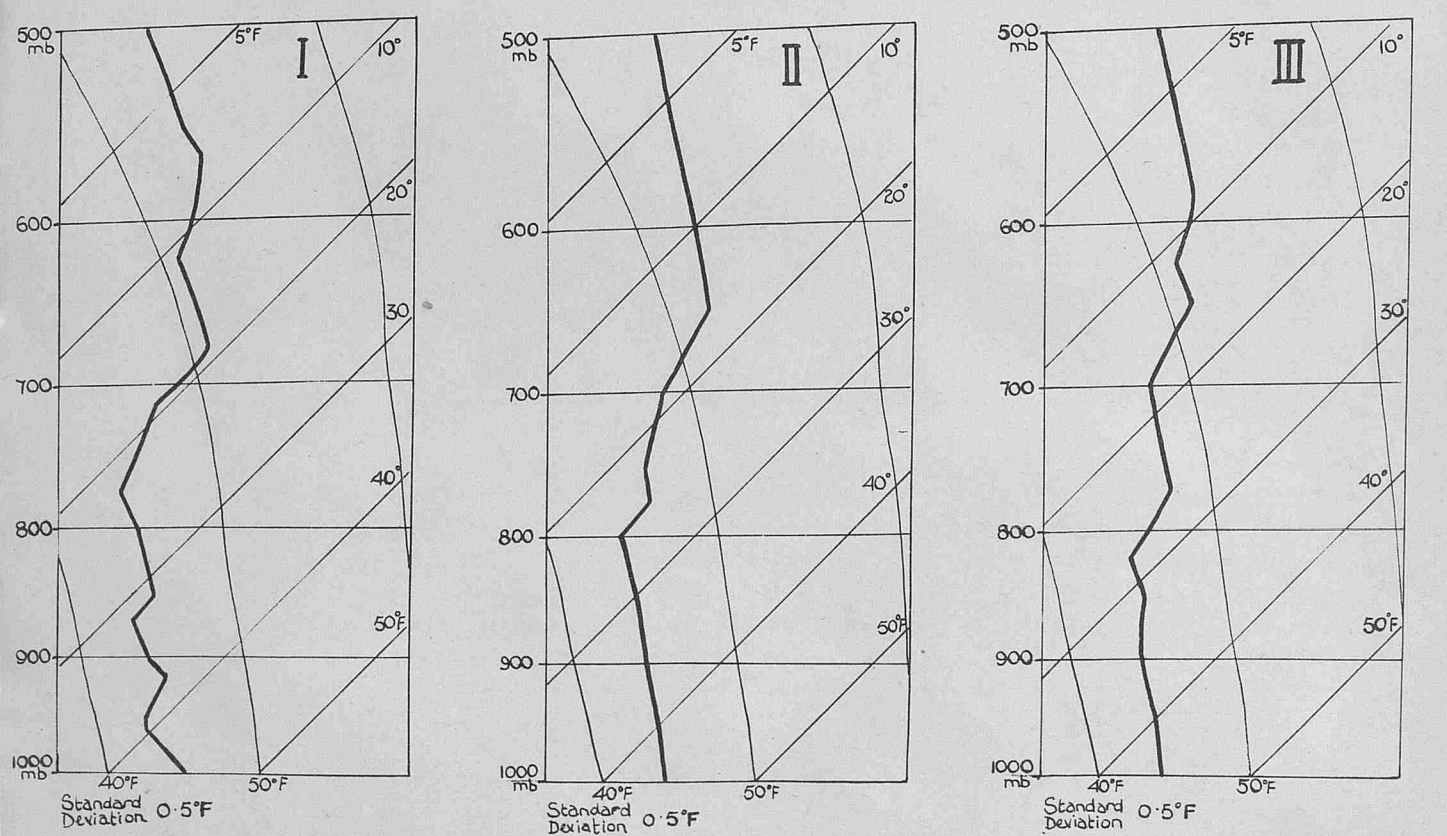


FIG 4(b) - Soundings through fronts "as reported"

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M.R.P. 776
S.C.II/133
24th November, 1952.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

The relation between wind flow over British Isles and the Mediterranean

by
C.S. Durst (Meteorological Office).

1. In the course of the investigations into the winds at 40,000 feet for Comet operations, correlation coefficients were calculated between the winds at 200 mb. over Larkhill and those over Rome and Malta. The particulars are shown in Table I and Table II, in which the components are resolved along and across a route running in the direction 325° to 145°.

TABLE I. Correlation coefficients between winds at Larkhill and at Malta 200 mb. (Dec., 1950 to Nov., 1951).

	Standard Deviations						Correlation Coefficients		
	Larkhill			Malta					
Season	Along	Across	SVD	Along	Across	SVD.	Along	Across	Stretch Vector
Mch.-May.	27	28	39	24	33	41	+0.08	-0.57	-0.30
June-Aug.	23	28	36	18	23	29	-0.12	-0.23	-0.20
Sept.-Nov.	33	32	46	25	30	39	-0.20	-0.37	-0.29
Dec.-Jan.	27	28	39	30	30	44	+0.03	-0.44	-0.22

Two observations a day were used so that for each quarter there were over 150 pairs of observations. Observations at 0300 made at Rome were also compared with Larkhill; many fewer were available, but between January, 1951 and October, 1951, 140 pairs were chosen and the result is given in Table II.

TABLE II. Correlation coefficients between winds at Larkhill and at Rome at 200 mb. (Jan. to Oct., 1951).

Standard Deviations						Correlation Coefficients		
Larkhill			Rome					
Along	Across	SVD	Along	Across	SVD.	Along	Across	Stretch Vector
26	28	38	20	25	32	0.07	-0.32	-0.15

/2.

2. From these tables it is clear that there is a significant negative correlation coefficient between the winds over the British Isles at 40,000 feet and those over the Mediterranean. The route which was being considered was one from London to Rome and if the westerly component of wind only had been considered the correlation coefficient might have been slightly greater.

3. Admittedly the correlation coefficients are not very high but even if we suppose that the values are of equivalent weight to only one fifth of the total number of observations the probable error of the coefficient for the W.S.W. components of Larkhill and Malta in spring would be 0.04, in winter and autumn 0.05 and even in summer only about 0.06.

4. I will not attempt to speculate on the significance of these figures in the mechanism of the general circulation. It would seem that the fluctuation in the Zonal flow in the great westerlies of Latitudes 25° to 40° may be partially compensated by the fluctuations in Latitudes 45° to 60°.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 777 by Mr. A.G. Forsdyke. A copy is available in the Library of the Meteorological Office."

M.R.P. 777
S.C.II/134
2nd December, 1952

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METEOROLOGICAL RESEARCH COMMITTEE

A study of depressions crossing the region of Labrador and the St. Lawrence Basin

By
A.G. Forsdyke (Meteorological Office)

1. Introduction.

One of the outstanding problems in extended forecasting for the British Isles is concerned with the behaviour of depressions over the northeastern regions of the American continent and the adjacent sea areas. It is well known that these systems have two broad modes of behaviour: they may either move out into the Atlantic or swing northwards and perhaps northwestwards to become slow moving somewhere to the west of Greenland. In the former case they may be expected sooner or later to bring cyclonic weather to the British Isles. The latter movement is frequently accompanied by at least the temporary building up of a ridge of high pressure on the Atlantic with the threat of a cold northerly type of weather over the British Isles. The present study was undertaken primarily to search for some criterion which would enable the broad mode of behaviour of particular depressions to be forecast long enough in advance to make it useful for medium-range forecasting. The associated large-scale thickness and upper flow patterns were the main basis of the study to begin with.

It was found that the period studied could be divided into spells each characterised by the similar behaviour of a few successive members of the sequence of depressions, and that each spell could be associated with some characteristic features of the hemispherical upper flow patterns. Little success was however achieved in the search for the medium-range forecasting criterion. It seems that the behaviour of these systems depends to a large extent on their detailed structure as well as the details in their own locality of the associated upper air patterns. There is considerable evidence for the thermal steering of these systems and, broadly in relation to the thickness patterns, the systems may be said to behave in accordance with the developmental ideas put forward by Sutcliffe (1947), but in most cases the surface and thermal patterns evolve together and give little advance indication for forecasting the behaviour of one on the basis of the other. But although its main object has not been achieved, this study has led to some interesting statistical results regarding the evolution of depressions of the type considered. The report is divided into two parts. Part I deals with the general behaviour of the systems studied and Part II consists of a statistical treatment, for selected groups of systems, of the relation between movement and deepening and certain features of the thickness pattern. Throughout this paper the term "contour" or "upper flow" refers to the 500 mb. contour pattern, and "thickness" to the 1000-500 mb. thickness pattern.

2. The data.

All depressions which occurred in a specified area (denoted by "L"), comprising Labrador and the northeastern part of U.S.A., during the year 1950 were selected for study. The boundaries of L are as follows:-

/To

To north, west and south, 60°N, 80°W, 40°N. respectively. To east and southeast, the lines joining the points 60°N, 50°W; 50°N, 50°W; 40°N, 60°W.

This area is indicated in Figs. 2 and 3.

The data were extracted from the 1:30 M circumpolar charts prepared twice daily at the Meteorological Office, Dunstable. All depressions which, while located in L, had at least one closed isobar (isobars at 4 mb. intervals) were considered. Track charts were plotted showing the position and central pressure of each depression at 12-hourly intervals over the period from two days before its centre entered L to three days after it left. The depressions were divided into two classes called "major" and "minor" respectively, according to their central pressure, as follows:

	"Summer" (April to September)	"Winter" (October to March)
Major	$\leqslant$ 1000 mb.	$\leqslant$ 992 mb.
Minor	$>$ 1000 mb.	$>$ 992 mb.

A depression was classified as "major" if for at least one position on the track chart the central pressure reached the value specified.

3. A general survey.

We begin with a general survey of the selected group of depressions the purpose being to provide the factual background for this study and an ordered assembly of the data. It is not suggested that the numerical values quoted have any special significance, for they are based on one year only and there is no reason to think that they would correspond at all closely with the data for other years.

The total number of depressions studied was 140, of which 106 (approximately 75%) were major systems.

3.1. Classifications of depressions according to movement.

The depressions were classified according to their movement into and out of L, the result being shown in Table I. For movement into the area the systems are classified as "W" if they entered across the western boundary, and so on; for movement out of the area the table is self explanatory. The plain entries refer to major systems, those in brackets to minor systems as defined in para. 2.

TABLE I

Frequency of depressions according to movement into and out of area L.

		Movement out of area L towards					
		W. Greenland Baffin Land N. Canada (Class N)	SE. Greenland Iceland (Class C)	SE. (Azores) (Class D)	Eastward into Atlantic (Class E)	Filled or absorbed by another low in or near L (Class F)	Totals
Formed in L		1 (1)	3 (-)	1 (1)	6 (1)	4 (5)	15 (8)
Movement into L from	W	25 (-)	13 (-)	1 (-)	12 (1)	14 (15)	65 (16)
	S	4 (1)	7 (-)	- (-)	10 (2)	3 (7)	24 (10)
	SE,E.	- (-)	- (-)	- (-)	- (-)	1 (-)	1 (-)
	N	- (-)	- (-)	- (-)	- (-)	1 (-)	1 (-)
Totals		30 (2)	23 (-)	2 (1)	28 (4)	23 (27)	106 (34)

/In

In regard to major systems only the main facts provided by Table I are as follows:-

(i) The majority of depressions originated outside L and moved into it from west (61%) or south (23%). The remainder nearly all formed within L; movement into L from north, east or southeast was very rare.

(ii) One half of the total number of systems moved out into the Atlantic, 27% more or less eastwards and 22% on tracks towards S.Greenland and Iceland. Only two systems moved definitely and consistently towards the Azores region.

(iii) Rather more than one quarter of the total number of systems remained west of Greenland after moving out of L. Of the 30 such systems only one had formed in L, four had moved into L from south and the remainder had moved in from west.

3.2. Monthly frequency of depressions.

For obvious reasons we are chiefly interested in the behaviour of depressions after they leave the area L. Accordingly for the purpose of this study they were divided into the following groups based on the classes of Table I.

Group N. Depressions which after leaving L remained west of Greenland.

Group C, D, E. Depressions which moved out into the Atlantic.

Group F. Depressions which filled or disappeared in L or near L (except Group N)

The monthly frequency of these groups is shown in Fig. 1. The distribution over the year is irregular and in no way remarkable. The occurrence of a large proportion (32 out of 50) of depressions of Group F during the summer months is in accordance with expectation Groups N and C, D, E. consist predominantly of major systems; of the total of 34 minor systems 27 filled in or near L.

3.3. Scatter of the depression centres.

The position of the centres of major depressions at intervals of from one to four days after they entered area L are shown in two series of charts, Figs. 2 and 3, (a) to (d). Fig. 2 is for systems which entered L from the south and Fig. 3 for those which entered from west. For convenience these two classes are denoted respectively by "S" and "W". In Figs. 2 and 3 the main seasons (Winter, January to March and October to December, and Summer, April to September) are distinguished by the type of symbol marking the position of the centre. For ease of identification on successive days the catalogue numbers used in the original tabulation of data for this study have been entered beside the position-symbols for some of the more interesting systems.

In a large number of cases the drawing of track charts for individual centres is not a simple, straightforward procedure. With few exceptions depression centres undergo complex processes of splitting, development of secondaries and absorption of other systems, so that in the drawing of track charts it is often necessary to exercise judgment regarding the identification of centres in successive positions.

Though Figs. 2 and 3 do not bring to light any systematic behaviour of the systems studied they do indicate certain broad features which are of some interest.

/(i)

(i) Of the 24 systems which entered L from south 16 (67%) were still in existence on the fourth day. Of systems entering L from west, 36 out of 65 (55%) were still in existence on the fourth day. The remainder had either filled, or been absorbed into other systems. Of those which filled, many had developed secondaries of one form or another and the secondary may in such cases be regarded as having taken over the role of the original system though the identity of centres is not preserved.

(ii) The Area to west of Greenland appears to be a specially favoured region for the filling of depressions, generally those of class W. (Only four systems of class S reached this area). It is well known that depressions moving from Labrador or Baffin Land often fill off W. Greenland; this filling is frequently accompanied by the formation of a secondary off SE. Greenland which moves away eastward.

(iii) The almost complete absence of centres from the mainland of Greenland is in accord with experience; it is however partly explained by the lack of observations. Only one of the systems studied (No. 27 Figs. 3(b), (c)) actually crossed Greenland. It is well known that depressions do sometimes appear to "jump" across Greenland from west to east; it is here suggested that these are systems which approach Greenland from west or northwest without traversing area L.

(iv) It is also well known that the area immediately to E. of the southern tip of Greenland is a very favoured location for depressions. This is not brought out by Figs. 2 and 3 so that we may conclude that depressions in this area are not, as a rule, those which have traversed L. It is believed that they frequently arise from the movements or developments referred to in (ii) and (iii) or are trough extensions from deep systems centred near Iceland.

(v) On the whole Figs. 2 and 3 confirm the well-known fact that depressions in middle and high latitudes move faster in winter than in summer. There are, however, one or two examples of very fast moving (late) "summer" systems, Nos. 104 and 107 (Fig. 2). These were both tropical storms, originating between Bermuda and Florida. The acceleration of tropical storms on passing into temperate latitudes has been referred to by Sawyer and Ilett (1950). No. 42 (Fig. 2), occurring in early April, only just qualifies as a "summer" system: its behaviour closely resembles that of the immediately preceding "winter" systems Nos. 40 and 41.

(vi) Little distinction can be drawn between the classes S and W as regards mobility. Figs. 2 and 3 tend to give the impression that the S depressions move eastward the more quickly of the two, but it should be pointed out that their starting point, as far as these diagrams are concerned, is farther east in all cases. It is apparent however that the S type depressions have little tendency to become slow moving on the western Atlantic, as the W type do as shown by the grouping S. of Greenland in Figs. 3(c), (d). The S type depressions also show a large scatter while there is a distinct tendency for W system to group themselves between 50°N and 60°N. up to the fourth day.

4. Types and spells associated with depressions in the Labrador - St. Lawrence region - general remarks.

A study of these depressions through the course of the year indicates that they tend to occur in groups or sequences consisting of about two to four members whose behaviour, particularly as regards movement, is similar.

/Such

Such a pattern of behaviour may persist for a few days or a week, occasionally longer. Its importance lies in the associated large scale pattern which may be affected over a sufficiently wide area to determine the broad pressure distribution over regions as distant as the British Isles. Table II lists the spells occurring during 1950 which lasted at least four days.

The dates given as the beginning and ending of a spell are respectively that on which the first members of the group of depressions comprising the spell entered L and that on which the last member of the group left L.

Seven types may be distinguished, based on the classes of Table I, with some sub-division. Table II give the characteristics of each type, the length of each spell and the total occurrence in days, in "summer" and "winter" respectively.

TABLE II. List of Types and Spells

Type No.	Description	Dates and duration of spell (Days)		Total No. of days Winter	Summer
N.1	Systems move N. or NW. bodily from area L to Greenland, Baffin Land or N. Canada. Little or no tendency for subsidiary systems to break away into the Atlantic.	Jan. 14 - 19	6	26	33
		May 6 - 16	11		
		June 10 - 19	10		
		Aug. 17 - 28	12		
		Nov. 20 - Dec. 5	16		
		Dec. 11 - 14	4		
N.2	Systems move as in Type N.1 but subsidiary systems break away into the Atlantic. The subsidiary system often extends SE. from the main centre; it frequently deepens while the parent system fills.	Feb. 18 - 23	6	19	14
		Mar. 23 - 27	5		
		July 11 - 15	5		
		Sept. 18 - 26	9		
		Dec. 24 - 31	8		
C	Systems move NE. from area L to E. of Greenland, sometimes beyond Iceland.	Feb. 24 - Mar. 4	9	24	12
		Mar. 28 - Apl. 12	16		
		Oct. 12 - 22	11		
E	Systems move in a general Ely direction from area L to mid-Atlantic, sometimes reaching the British Isles.	Feb. 4 - 17	14	23	18
		Mar. 12 - 20	9		
		June 20 - 26	7		
		July 23 - Aug. 2	11		
CE	Combination of C and E. Either, systems of types C and E alternate, or they move as type E to mid Atlantic and then turn sharply N. towards Iceland.	Jan. 2 - 12	11	49	40
		Jan. 26 - Feb. 3	9		
		Apr. 23 - May 3	11		
		Aug. 9 - 17	9		
		Aug. 31 - Sept. 18	19		
		Sept. 30 - Oct. 6	7		
		Nov. 5 - 18	14		
		Dec. 16 - 24	9		
F	Systems fill in area L, or within a short time after leaving L.	Apr. 12 - 23	12	-	34
		June 3 - 8	6		
		June 27 - July 8	12		
		July 17 - 20	4		
H	No depressions in area L, which is often anticyclonic.	Jan. 20 - 25	6	14	20
		May 20 - June 2	14		
		Aug. 3 - 8	6		
		Oct. 27 - 30	4		
		Dec. 6 - 9	4		

/Detailed

Detailed studies were made of these spells primarily with object of discovering any common features in the large scale surface and upper air patterns which would enable the onset of a spell to be recognised beforehand. Such common features as there are, however, are recognisable only on a very broad basis and differ so much among themselves in small and medium scale features that their value for forecasting cannot be very great.

The results of these studies are discussed in paras. 5 to 10. Lest it should seem that an undue amount of attention is given to type N1 we may recall that the broad distinction we have adopted lies between stagnating systems (N1) on the one hand and mobile systems (N2, C, E) on the other.

5. Type N1. Systems moving bodily from Area L to Greenland, Baffin Land or Hudson Bay with only a weak tendency to break away E., NE., or SE. into the Atlantic.

5.1. General.

These systems though not associated with significant forward break-aways were, as a rule, complicated by secondaries moving from southwest; these secondaries usually became absorbed into the main system and so are classified as "F" in Table I. In five out of the six spells also, the original depression was the first of a group of two or more independent systems moving from southwest each intensifying in turn near L and taking over the role of the main system, defining the type. Such renewal or replacement would seem to be necessary to the establishment of a spell, for in the five cases of single depression of this class the centre did not survive as the main feature in the vicinity for the necessary four days, and no spell was recorded.

The northward or eastward movement of the centre was in all cases broadly consistent with the direction of the thermal steering, implying a thermal trough near the longitude of Hudson Bay. In most cases (14 out of the 18 studied in detail) this trough was already in existence in high latitudes before the beginning of the spell, with frequently a cold pool over Hudson Bay or the Canadian Archipeligo. Sometimes a thermal ridge developed northwestwards from S. Greenland across Baffin Land to the extreme north of Canada, separating the Hudson Bay cold pool from the polar basin. The southward extension of the cold trough, necessary for northward steering from area L, was, however, most often effected in conjunction with the low itself over area L, rapidly transforming an eastward or northeastward steering pattern into a northward or northwestward one. It appears therefore that the pre-existing thermal pattern over L affords little guide to the onset of a spell of this type. We return to this point in more detail in para. 11.

5.2. Large scale patterns associated with type N1.

The definition of this type implies some degree of stagnation in the vicinity of Labrador: if there is any coherence at all in the large scale surface patterns it would be reasonable to look for immobility in other parts of the hemisphere also. A study of the hemispherical 500 mb. flow patterns during the six spells listed in Table II does indicate that there is immobility of the large scale troughs and ridges in a broad sense, though the patterns are disturbed from time to time by small mobile features. Characteristically also the troughs and ridges are features of large amplitude, and usually they persist through the whole of a spell. But though the six spells have these broad features in common, there are important differences between them, arising from the longitudinal locations of the ridges and troughs, (apart from the trough over eastern North America, which, from the type definition, is to be expected as a constant feature). The sectors of the hemisphere which were occupied by slow moving and persistent upper troughs and ridges in the six spells are shown in Fig. 4. Troughs are marked by full lines and ridges by broken lines drawn along arbitrarily selected parallels of latitude, one for each spell. For easy reference

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the spells are numbered in Fig. 4 and the corresponding dates (from Table II) are also given. Generally the extent of the sector marked indicates the range over which the axis of the trough or ridge varied during the spell, but for period 1 the line extending from 130°W. to 40°W. indicates that the whole of this sector was occupied by a broad trough. On the whole the axes of the troughs and ridges were inclined at only a small angle to the meridians so that little difficulty arises in placing a trough owing to the latitudinal variation of the position of its axis. The overlapping lines in the Pacific for period 4 do however indicate a quasi-stationary "cyclonically" distorted pattern consisting of a ridge in high latitudes and a trough farther south with their axes lying from NW. to SE.

Fig. 4 shows clearly that the locations of the persistent upper troughs and ridges differ between spells. In particular there is a marked shift to westward of the ridge-trough pattern over the Atlantic and Europe in the last three as compared with the first three spells. It is evident from the dates of these spells that this shift is not a seasonal effect, as might be expected having regard to the fact that the mean zonal flow and the associated spacing in the long wave pattern are greater in winter than in summer. Other features which are common to most of the spells, though still somewhat variable in location are the ridge near the E. Pacific, the trough in the east-central Pacific and the trough off the E. Asiatic coast. No features are shown over the eastern half of Asia because of the scarcity and uncertainty of the observations there.

A further characteristic of the upper flow patterns associated with this type is that in high latitudes the circumpolar vortex is usually split up into a number of separate centres each covering a substantial area. One of the most prominent of these lies over Labrador or Baffin Land, separated from the polar basin by a ridge across the Canadian Archipeligo from the NW. Atlantic to Alaska.

Associated with the persistent upper troughs and ridges are large-scale cyclonic and anticyclonic circulations at the surface - often with their longer axes and orientated more or less meridionally. The individual cyclonic regions remain on the whole distinct and separate throughout a spell, with little or no tendency for depressions, except very weak ones, to pass from one to the other. But as in the case of the upper troughs and ridges there were no geographical regions (apart from Labrador) which preserved consistently cyclonic or anticyclonic regimes throughout all the spells. In one spell (June 10th - 19th) the surface patterns were quite disorderly consisting of numerous small and weak systems which showed little constancy from day to day. Of the remaining five spells, four were characterised by persistent cyclonic activity in the region Bering Strait - Gulf of Alaska - E. Pacific, three in the region British Isles - Faeroes - Scandinavia and two in the region W. Pacific - NE. Asia. Perhaps more characteristic of these five spells were the persistent meridional anticyclones extending from the subtropics to the polar basin, though these differed somewhat in orientation and location from spell to spell. The last three spells had persistent meridional anticyclones over Greenland and mid-Atlantic. In other spells and in other sectors, meridional highs, though not persisting throughout the whole period were sufficiently prominent and lasting to set the general character of the surface circulation.

Near the British Isles the broad features of the surface patterns in the six spells were as follows:-

January 14 - 19th.	Straight isobars between high to N. and low to S. at first; later anticyclonic.
May 6 - 16th.	Anticyclonic easterly or northerly throughout: high centre shifted from N. of British Isles to Greenland.

/June

June 10 - 19th.	Pattern complex with numerous small weak systems.
August 17 - 28th.	Cyclonic southwesterly.
November 20th-December 4th.	Cyclonic northerly.
December 11th - 14th.	

By way of illustration we now proceed to a more detailed discussion of two of these spells, chosen from the summer and winter months respectively.

5.3. The spell of 17th - 28th August, 1950 (Type N1).

The surface charts for four days spaced as nearly as possible at equal intervals over the spell are reproduced in Figs. 5(a) to (d). The regular time interval was the only consideration leading to this particular selection of charts; the intermediate charts do of course show some variation from these but the broad features of the large scale type are to be found on almost all the daily charts throughout the spell.

The surface charts show a persistent cyclonic region over Labrador, Hudson Bay and Baffin Land. Five separate depressions passed across area L during the spell, four of them becoming slow moving in the way characteristic of Type N1. The other depression originated as a tropical storm, moving into L from SW. and passing away E. into the Atlantic. This interruption, which lasted for two days only, (22nd and 23rd August) was not regarded as important enough to break the spell.

There are examples of persistence also in some other parts of the hemisphere, apart from the low pressure over the southern half of Asia and the high pressure over the sub-tropical Atlantic and Pacific which are seasonal or semi-permanent features. Persistent cyclonic areas are (i) Bering Strait, Alaska and the Aleutians (ii) the British Isles and NE. Atlantic and (iii) the Urals. Anticyclonic situations persist over (i) Greenland and the western N. Atlantic, (ii) N., Central and SE. Europe and (iii) NW. Canada. There is probably sufficient persistence in these features to specify a large-scale circulation type, which, however, depends more upon the general lack of mobility than on the precise geographical location of synoptic systems. Furthermore, considerable changes are to be seen in some areas; for example the western half of N. America is anticyclonic in the first part of the period and cyclonic in the second. On the whole the surface circulations have a meridional character, as is to be expected from the large amplitude features of the 500 mb. pattern, shown in Figs. 6(a) to (d). In these charts the axes of troughs and ridges have been marked for persistent features only. There are persistent troughs over (i) the E. Pacific (ii) eastern America and the western Atlantic, (iii) the eastern Atlantic and the western seaboard of Europe and (iv) the Black Sea and Caspian area. Ridges persist over (i) western America and Alaska (ii) the west-central Atlantic and (iii) central and northern Europe. The several centres of the circumpolar vortex are well shown on all four charts.

The day to day persistence of the 500 mb. troughs and ridges is illustrated by Figs. 7(a) to (c). These are composite charts each showing the configuration of a selected contour (18,600 feet) on four successive days, the three charts together covering the whole spell. The persistent troughs and ridges of Fig. 6 will be easily recognised throughout the series and it will be readily seen that the day to day changes are small. The major troughs are particularly constant over the last four days of the spell (Fig. 7 (c)); the fact that the greatest degree of steadiness in the upper flow patterns is reached at this period suggests that the general pattern of the spell is set more by the circulation at the surface than at the upper levels.

/5.4.

5.4. The spell of 20th November to 5th December, 1950 (Type N1).

This spell is illustrated by Figs. 8, 9 and 10 which show the surface, contour and composite charts respectively as in para. 5.3. Fig. 8 shows that the eastern and northeastern parts of the U.S.A. and Canada were persistently cyclonic. During the period eight separate cyclonic systems traversed the area L. Four of these were short-lived systems which filled within L; the remaining four systems became slow moving to northward of L thereby determining the type. The most notable of the other persistent features was a large complex cyclonic area near the British Isles and extending NE. across Scandinavia to beyond Nova Zemlya; at times this complex covered the Faeroes and Iceland also; some of its individual centres became very intense. This region of cyclonic activity was at times linked with a weaker one which persisted near the Azores, but for part of the period these cyclonic areas were separated by a pronounced ridge. Cyclonic activity, rather weak on the whole, persisted in middle and low latitudes on the E. Pacific, while Alaska was mostly anticyclonic. The Asiatic anticyclone was well established and in the main there was high pressure over N. America; these however were seasonal features and not especially connected with the broad type. On the other hand, the persistent anticyclones over the central Pacific and the west-central Atlantic appear to have been important features of this particular spell, and were effective throughout in separating the main cyclonic areas.

Corresponding to these surface features there are shown in Figs. 9 (a) to (d) persistent large scale troughs in the 500 mb. pattern, situated over (i) E.Asia and the W. Pacific (ii) the E. Pacific (iii) eastern America and (iv) western Europe. There are persistent ridges in mid-Pacific, mid Atlantic and over eastern Europe, that is between three of the trough-pairs, but no pronounced ridge appears in the American sector. The composite charts of Fig. 10 show that these main features were on the whole persistent. The troughs over eastern America and western Europe became complex at times and with the intervening ridge were subject to considerable distortion about the middle of the period. This distortion was associated with a good example of "double structure" in the surface pattern from America to western Europe (Figs. 8(b) and (c)), with two regions of cyclonic activity in high and low latitudes respectively separated by a zonal ridge. As in the period discussed in para. 5.3 we again notice that the upper patterns tend to become steadiest towards the end of the spell.

6. Type N2. Systems traversing area L move away towards the north as in N1 but subsidiary systems break away into the Atlantic. The original centre usually fills and the subsidiary often deepens to become a major system.

6.1. General.

The subsidiary systems appear either as warm front waves (Sawyer, 1949) or develop at the point of occlusion of the main system. In the latter event they usually originate as small waves moving from SW. on the main cold front. The most marked development seems to occur when the thermal trough from Davis Strait penetrates far SE. into the Atlantic: cyclonic development is then propagated southeastwards in association with the tip of the thermal trough. The details of the development however appear to depend on the particular circumstances of each case. With cyclonic systems breaking out into the Atlantic from the west any meridional build up of high pressure in the longitude of Greenland is likely to be short lived. A latitudinal belt of high pressure may, however, be a persistent feature, usually in the normal low-latitude position, but occasionally in middle and high latitudes with the cyclonic activity coming out on a track well to the south.

/6.2.

6.2. Large scale patterns associated with Type N2.

In so far as the depressions traversing area L stagnate to westward of Greenland, this type might be expected to show the slow moving features characteristic of Type N1. On the other hand the breaking away of substantial systems to E. or SE. suggests some degree of mobility in the patterns. An examination of the 500 mb. contour charts confirms these expectations to a considerable degree. In all five spells there are present a number of slow moving troughs of large amplitude. These troughs are on the whole less well defined than in Type N1 while the intervening ridges are much less conspicuous. Between the troughs the patterns are often zonal in the middle and lower latitudes, clearly associated with the tendency for subsidiary depressions to break away eastward from the slow moving primaries. In the higher latitudes the circumpolar vortex is usually broken up into a number of distinct centres, from which the lower latitude troughs extend more or less meridionally. Between these centres the flow patterns are weak and irregular with few well-defined ridges.

The main troughs are often persistent though rather more mobile than in Type N1. Fig. 11, corresponding to Fig. 4, shows the sectors occupied by persistent large scale ridges and troughs in the five spells of Type N2. Again we see differences between the spells but as in Fig. 4 the troughs over the W. Atlantic and the E. and W. Pacific are present on most occasions. The shift of the European trough from Russia in the earlier part of the year to western Europe in the latter part, noted in Fig. 4 is seen also in Fig. 11. An examination of the daily charts shows that as compared with Type N1 the upper troughs in Type N2 tend to be broader, with strong nearly straight currents on their southern peripheries. Small features run along these currents and often continue more or less eastwards in the low and middle latitude zonal flow between the troughs.

In regard to surface features, the only areas of the hemisphere in which conditions were persistently similar in all five spells were the subtropical oceans where the normal anticyclones were more than usually elongated in the west to east direction. The polar basin was on the whole anticyclonic, though not entirely so throughout any spell. In four of the five spells there was a persistent cyclonic situation in the Gulf of Alaska. Otherwise the surface situation showed gradual changes owing to both movement and development in each of the spells.

6.3. The spell of 18th - 26th September, 1950 (Type N2).

This spell is illustrated by the surface and 500 mb. charts for three equally spaced days (Figs. 12 and 13) and by composite charts of the 18,600 feet contour (Fig. 14). During this period two major depressions entered area L from the region of Hudson Bay, and swung northward or northwestward; both had prominent subsidiary systems which broke away into the Atlantic. Fig. 12(b) shows the first primary near Hudson Bay with its breakaway system south of Greenland. In Fig. 12(c) the second primary is in Davis Strait and the incipient breakaway appears as a trough to southeast of Greenland. Apart from area L and its neighbourhood, other persistently cyclonic areas during this spell were (i) NW. Canada and Alaska and (ii) Scandinavia and northern Eurasia; the cyclonic activity in (ii) extended at times over the British Isles. The subtropical high pressure belt persisted right round the hemisphere apart from a few small gaps. The contour patterns of Fig. 13 show the persistent troughs over the E. Pacific, the W. Atlantic and Europe and the ridge over America. These features are less well defined than in Type N1. The nearly straight flow in middle latitudes is well shown for the Atlantic in Fig. 13(a) and the Pacific in Fig. 13(b). The persistence or otherwise of these features of the flow pattern appears from the composite charts; the large scale troughs are persistent on the whole but there are numerous small mobile features. We may notice that in this example also the flow pattern does not reach its most steady state until the latter part of the spell.

/7.1.

7.1. Type C. Depressions moving NE from area L towards and often beyond Iceland.

In view of the widely recognised association of depressions with the Iceland - SE. Greenland area the small number of spells with depressions moving from L into this area seems a little surprising; this has been remarked upon in para. 3.3. Also, some of the depressions which enter this area are interspersed with systems more directly eastwards and are included in the combined Type CE of Table II.

On the whole the movement of depressions of Type C conforms to the direction of thermal steering. Generally there is a large scale thermal trough over the Hudson Bay region; the trough is widely open to northwards and has a strong thermal gradient on its southeastern flank. A common feature of the four spells recorded is the "cold pool" near NW. Greenland, i.e. farther east than in Types N1 and N2, which is in conformity with northeastward rather than northward steering. This location of the cold pole in association with depressions in the Iceland area appears to agree with a recent (unpublished) finding of Brezowsky (1952) that there is a relation between the position of the cold pole over Canada and the character of the European winter: if the cold pole is west of Hudson Bay winters in Europe tend to be cold and anticyclonic, but if to the east of Hudson Bay European winters are milder and more cyclonic.

7.2. Large scale patterns associated with Type C.

The 500 mb. flowpatterns associated with this type are more complex and show a greater tendency to change from day to day than in Types N1 and N2 already discussed. They are characterised by numerous troughs and ridges in the middle and lower latitudes. In the higher latitudes the patterns are somewhat simpler and usually show a large and compact polar vortex (i.e. one without narrow straggling troughs or numerous small centres) having a kind of two-fold symmetry with lobes in the diametrically opposite sectors W. Atlantic and W. Pacific. The troughs and ridges of the inner and outer parts of the pattern are often in antiphase, giving rise to well marked confluences and difluences, and to localised bands of strong flow (jet streams). In the three spells studied there were two persistent troughs, one over E. America and the W. Atlantic and the other over Europe. Otherwise the upper troughs and ridges were temporary and mobile features though often moving rather slowly.

This is essentially a mobile type and there are few persistent stationary features in the surface patterns. The depressions moving into the Iceland region from the SW. often deepen there and contribute to the formation of a large cyclonic complex which extends at times over much of NW. and N. Europe, Iceland and Spitzbergen. Intermittent weak cyclonic activity over SE. Europe and the Mediterranean is associated with the persistent upper trough over Europe mentioned above: there is often some steering of cyclonic activity from high latitudes southeastward on the western flank of this trough. The Aleutian low is a common but not a persistent feature. The Siberian high is usually extensive, with its main centre rather towards the east of the land mass. The polar basin is often in a col between cyclonic regions over the N. Pacific and extreme NE. Atlantic and anticyclones over Canada and E. Siberia; the col may be cyclonic or anticyclonic its character varying within a given spell.

7.3. The spell of 24th February to 4th March, 1950 (Type C).

Charts for this spell are shown in Figs. 15, 16, 17. We are here interested in illustrating a type in which individual systems are mobile and therefore select three days at short intervals on which the movement of one and the same depression can be followed. This is the depression near the Great Lakes on 1st March (Fig. 15(a)), over Labrador on 3rd and off NW. Norway on 5th; it is the same as No. 32 in Fig.3. The variable nature of the large-scale surface patterns is illustrated by the changes

/over

over Canada, the central part of northern Asia and the polar basin. The 500 mb. patterns are zonal in character, with the main flow well out from the pole in most sectors. There are several examples of high and low latitude ridges and troughs in antiphase. The composite charts for the first eight days of the spell (Fig. 17) show that the large scale characteristics are preserved: on the whole the flow patterns are smooth, with small mobile ridges and troughs. The more permanent troughs on the W. Atlantic and over Europe are common to all three spells of this type.

8.1. Type E. Depressions moving in a generally easterly direction from L to mid-Atlantic, sometimes reaching the British Isles.

In all four spells the general eastward movement is associated with a zone of thermal concentration running roughly eastwards through L and well out into the Atlantic; it is generally well marked in winter, but weaker and variable in summer. In the two winter spells the thermal concentration lies on the southern flank of a large cyclonic thermal eddy which has its root near Hudson Bay and protrudes far into the Atlantic. Considerable cyclonic activity is associated with the exit to the thermal "jet" thus formed.

In two spells, March 13th - 20th, and June 20th - 26th, 1950 there was a persistent thermal ridge with an associated anticyclone over Greenland. This thermal pattern resembles that which occurs with Type M1, but is more twisted. In Type M1 the ridges are blocking features of large meridional extent; in Type E they are high latitude features undercut by the cyclonic features moving east in middle and lower latitudes. In the spell from February 4th - 17th there was also high pressure over Greenland for the most part but it was associated with an overlying thermal trough.

8.2. Large scale patterns associated with Type E.

At the 500 mb. level the inner part of the circumpolar vortex is complex, with many small centres and weak gradients. The main westerly flow is well out from the pole but is often much more distorted than with Type C. The main upper troughs are broad features, flanked to the south by nearly straight belts of strong upper flow; depressions move quickly eastward in these belts and, swinging to the left, deepen and become slow moving in the weak upper flow to the north. Between the broad troughs there are rather narrow and often distorted ridges. The troughs are often persistent but the ridges are transitory and subject to rapid changes of amplitude.

The large scale surface patterns show marked seasonal differences. By definition, the type is associated with much cyclonic activity in the N. Atlantic. In winter this fits naturally into the climatological mean picture, together with a similar large and active cyclonic area over the N. Pacific and the continental anticyclones of Siberia and N. America. This suggests that this type of winter circulation is a stationary mode, with the thermal effects of the undersurface and the dynamically balanced strong zonal flow with long wave-spacing, mutually in phase. In summer the type requirement (by definition) of oceanic low pressure (at least over the N. Atlantic) having presumably a dynamical cause, would appear to conflict with the thermally induced monsoonal low pressure of the continents. The resulting surface patterns are rather ill-defined with many small and weak centres; only the subtropical anticyclones appear as well formed and lasting features.

8.3. The spell of March 12th - 20th, 1952 (Type E).

Surface and 500 mb. charts for this spell are given in Figs. 18 and 19 at four day intervals. The basis of classification for this period is the large cyclonic area on the Atlantic, which the daily charts show to have developed mainly from depressions moving eastward from area L. The centre of this cyclonic complex is almost ten degrees south of the

/normal

normal seasonal position for the Icelandic depression. There is a similar cyclonic complex in the Pacific. It is interesting to notice that the two systems have somewhat similar histories - an eastward movement and intensification followed by a splitting into two large components the axis of the composite system lying SW.-NE. A similarity appears also in the 500 mb. patterns which, to begin with, have marked cyclonic distortions in the Atlantic and Pacific. The subsequent degeneration of these features is seen in Figs. 19(b) and (c). The composite chart of Fig. 20(a) shows that the Pacific and Atlantic ridges disappear simultaneously and in a similar way. Subsequently the flow pattern settles down to a steady smooth type in most sectors, apart from some mobile medium amplitude features over Europe and Asia.

9. Other types.

The remaining types CE, F and H were also studied, but are only briefly discussed in this paper for the reasons given in paras. 9.1 and 9.3.

9.1. Type CE. Spells combining the characteristics of types C and E.

This, a combination of two fairly common types, occurs the most frequently of all the types considered. Individual lows from area L move into the Atlantic but fall into no particular grouping; movements towards NE., E and SE. appear to occur more or less at random. Some systems combine the characteristics of types C and E by moving more or less directly eastward to mid-Atlantic and then turning sharply northwards towards Iceland.

The large scale patterns have some characteristics of both the preceding types, discussed in paras. 7.3 and 8.3, so that they will not be dealt with in further detail here. The thermal trough over east America is usually broad in shape and extends out over the W. Atlantic. There is also a broad trough at the 500 mb. level, which tends to be somewhat to eastward of the thermal trough; this is associated with a semi-permanent surface depression near or somewhat to eastward of Iceland. Systems moving out from N. America feed into or renew this semi-permanent feature from time to time. The steering of these systems appears to conform more nearly to the contour than to the thickness pattern.

9.2. Type F. Depressions filling in or near area L, soon after leaving it.

Table II shows that spells of this type in 1950 occurred in the summer half year only. There are no common and easily recognisable characteristics of the thickness and contour patterns associated with it. On the whole the thermal gradients are weak and irregular. The vicinity of area L, where depressions are filling, is an inactive region and would hardly be expected to have appreciable influence on the circulation patterns of a wider area.

9.3. Type H. No depressions in area L, which is often anticyclonic.

These include all spells in which no actual low pressure centres were situated in area L. In most cases the area lay in the region of "straight isobars" between high and low pressure regions. On a few occasions however the area was actually covered by an anticyclone. Four out of the five spells recorded lasted only six days or less; the remaining spell, from May 20th to June 2nd, lasted 14 days. An examination of the daily charts during this spell shows that conditions were by no means persistently anticylonic in area L which was at times invaded by weak frontal troughs. The upper patterns showed a tendency for the presence of a ridge in the Hudson Bay region (where the mean thickness charts show a trough). This ridge however was a recurrent rather than a persistent feature, and it was interspersed with irregular upper patterns.

/10.

10. Changes of type.

Considered in relation to the problem of medium range forecasting, the onset and breakdowns of these spells, being associated with type changes, are equally as important as their persistence. We naturally look to the flow pattern in its broader aspects for any possible advance indication that the flow is about to settle down into a quasi-stationary state such as accompanies certain of these spells. Bearing in mind that from the large scale point of view the most easily recognisable distinction lies between the stagnating type M1 and the remaining mobile types, a special study was made on the onset and breakdown of this one type. The hemispherical upper flow patterns were examined over at least the four day periods and following each spell as listed in Table II. The result of this study was not sufficiently encouraging to extend it to other types, and instead of the definite forecasting criteria which, it was hoped, would emerge we can make only a few rather uncertain suggestions.

10.1. Onset of Spells (Type M1).

It was stated in para. 5.1. that depressions of Type M1 are associated with a thermal trough near Hudson Bay, but that before the beginning of the spell the thermal trough is in evidence only in the higher latitudes. We have also seen that there is usually a large amplitude trough at 500 mb. in the same longitude after the M1 character of the depression is established. It is obviously important to investigate the question as to whether the upper trough exists beforehand in which case the movement of the depression is brought about by simple steering in the upper current, or whether the depression itself is a necessary factor in establishing the upper trough largely by thermal advection.

The surface and 500 mb. charts were examined over a period of four days covering the onset of each M1 spell. In each case there was an upper vortex in high latitudes over or to the westward of Baffin Land before the spell began. But in five out of the six spells the southward extension of the trough to form the usual large-amplitude feature was not effected until the depression had, so to speak, moved into position east of the growing trough. In the remaining case the upper pattern was in existence before the spell began; in this instance (Dec.11-14th), however, the spell followed within six days of the end of a similar spell and the previously existing upper pattern was scarcely disturbed. The interval between the two spells was marked by an intense anticyclone in the vicinity of L, which maintained the large amplitude upper patterns except for a slight shift of phase.

These changes accompanying the onset of a spell are illustrated in Fig. 21, which shows the 500 mb. patterns for the four days 19th - 22nd November, 1950 and the first and second members of the depression sequence. It is clear that the large-amplitude trough is not established until the 21st or even the 22nd November and that the circulation of the depression is a principal agency in its development.

It is sometimes argued that the formation of such a large-amplitude upper trough and depression complex is a purely dynamical effect, that is, a downstream wave in the upper flow pattern resulting from barotropic flow in association with a large amplitude upstream feature. On this basis, the upper trough would be causal and the surface depression consequential. The characteristic large-amplitude quasi-stationary upper flow patterns which we have found to be associated with the M1-type lend some support to this argument. On the other hand we have just seen that the upper trough over eastern America does not reach its characteristic full development until after the depression has moved into a suitable position. This would suggest that, though dynamical considerations would require a large scale upper trough in particular sector, it will not form unless there is a depression already in existence which can phase-in with the developing upper trough; that is the surface and upper patterns must be regarded as playing equally important roles in this development.

/Having

Having regard to the large-scale patterns it seems reasonable to look for signs of the development of the characteristic large amplitude slow moving ridges and troughs in other sectors of the hemisphere as well as that which we are primarily concerned, namely, eastern America and the Atlantic. In particular we would study the developments in the sector upstream of this one, with a view to judging whether the idea of downstream propagation of large scale features of the upper flow can be useful. We consider the six N1 spells in some detail from this point of view.

10.11. January 14th - 19th.

The preceding type was CE but, except in the E. America-Atlantic sector, the large amplitude upper flow features characteristic of Type N1 were present several days before the spell began. Over the Pacific such a pattern had been in existence from January 6th. From 120°W to 40°W the main 500 mb. flow was almost "straight" on the southern flank of a very broad trough. On January 11th - 12th an additional large amplitude trough-ridge appeared temporarily in the contour and thickness patterns in mid-Pacific, suggesting an unstable oscillation, for it is difficult to explain the new feature on advectional or dynamical grounds. On January 16th there occurred the rapid building of a "blocking" anticyclone near E. Greenland in a manner which seems inexplicable on thermal development ideas. Though clearly not connected with the onset of the spell this occurrence was an important factor in its maintenance.

10.12. May 6th - 16th, 1950.

The preceding type was CE; up to May 1st it was accompanied by a smooth flow pattern with only small-amplitude perturbations round most of the hemisphere. During the period 3rd - 5th May complexities with cut off cold pools and distorted ridges and troughs appeared simultaneously over the Rockies and Europe. Over W. Europe there developed a marked oscillation of short wave-length accompanied by the building of a blocking anticyclone over Scandinavia and the British Isles. During the first few days of the spell this anticyclone and its associated thermal ridge spread westwards into the Atlantic and blocked the eastward movement of depression from area L.

Between May 5th and 8th a depression moved steadily NE. across the U.S.A. from Mexico to Labrador and became a very large and deep system. The intensity of its development is difficult to explain on the basis of the thermal pattern. Becoming slow moving in Davis Strait from May 8th - 10th this system was apparently the cause of very large changes in the upper patterns, building a ridge in the Atlantic where hitherto there had been a marked trough, and leading to general retrogression in that sector.

10.13. June 10th - 19th, 1950.

The preceding type was F. About a week before the beginning of the spell there existed a quasi-stationary pattern which, except for a longitudinal displacement, would have been classified as N1. Slow moving cyclonic systems were situated over the E. Atlantic, N. Russia and to the W. of Hudson Bay with associated large amplitude upper troughs and ridges. From June 6th to 9th the sector from mid-Pacific to the Great Lakes was occupied by a mobile sequence of fairly large amplitude troughs and ridges while a smooth pattern existed in the Atlantic. During the next four days mobility was transferred to the Atlantic while over the Pacific and W. America the flow gradually settled down to the large amplitude quasi-stationary type. Later these characteristics appeared in the Atlantic also. This spell therefore affords some evidence for the downstream propagation of flow pattern characteristics.

/10.14.

10.14. August 17th - 28th, 1950.

The preceding type was CE. Between August 13th and 16th large amplitude oscillations were already present in the upper flow patterns over the central Pacific and E. Europe, but in the Atlantic the pattern was rather flat. The subsequent appearance of the characteristic oscillation in the Atlantic appears to have been effected mainly by retrogression of the European trough-ridge system. This was aided by the formation on August 19th and 20th of a blocking anticyclone on the W. Atlantic which, since it developed forward of a difluent thermal trough and extended well northwards, is difficult to explain on the basis of thermal development.

10.15. November 20th - December 5th, 1950.

The preceding type was CE. Apart from a well-marked ridge in the central Pacific the pattern was generally flat on November 16th, and the Pacific ridge itself soon flattened also. But on November 17th - 18th a large amplitude ridge development over E. Europe as an intense anticyclone formed near Nova Zemlya: a day or two later a blocking anticyclone developed near Greenland with a pronounced upper ridge on its western flank. There is a suggestion here of the upstream propagation of blocking action.

10.16. December 11th - 14th, 1950.

The preceding type was H. Large amplitude features were present in the upper flow in most sectors of the hemisphere from December 7th - 10th, and these were, on the whole, mobile. On December 11th and 12th a blocking anticyclone built from the Azores to Greenland, while an upper trough in mid Atlantic was replaced by a ridge. The main new anticyclonic development occurred over Greenland itself, in a "natural" position to the left of a thermal confluence.

10.2. Summary of conditions preceding the onset of spells (Type N1).

It is clear that the conditions antecedent to the spells discussed in 10.11 to 10.16 have little in common that could be used as an aid to extended forecasting. There is no consistent evidence of either downstream or upstream propagation of the large amplitude characteristics of the upper patterns. In this connexion the spells might be classified as follows:

Downstream propagation 2 periods (10.11, 10.13).

Upstream propagation 1 period (10.15).

The remaining three cases were indefinite. In three of the periods there was development of important surface features which would not have been anticipated on thermal development ideas. These were blocking anticyclones in the Atlantic in two spells (10.11 and 10.14) and a deep depression over America (10.12). An "unexplained" upper oscillation occurred in the Pacific in the January spell (10.11). These suggest that the type changes are brought about by the setting in of instability in the upper or surface patterns, possibly both together. In any case no precursory signs of the change were detected. In two of the periods (10.12 and 10.14) there was westward movement of an important system from W. Europe to the Atlantic (retrogression with blocking).

10.3. Breakdown of spells (Type N1).

Of the six spells, three were followed by type H (anticyclonic in L) and three by type CE or E (depressions moving NE. or E. from L into the Atlantic).

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Of the spells followed by type H, two occurred in winter (January and November-December). The anticyclone in each case was very large and intense, its existence doubtless being helped by the cold land mass to westward of L. Dynamically the upper patterns were not inconsistent with anticyclogenesis, though at the same time not appearing to demand it. The advection thrust cold air well southwards on the western Atlantic and in effect displaced the upper trough eastwards from the Hudson Bay region; though this effect was to some extent masked by the seasonal trough over the continent. In both these cases the large amplitude troughs and ridges were preserved beyond the end of the spell, but became mobile. The remaining spell followed by Type H occurred in May; the anticyclone in area L was one of many small features in a very complex and distorted general pattern, and its occurrence in that particular locality seems (on the broad scale) to have been largely fortuitous.

The breakdown of each of the remaining three spells was preceded by a degeneration of the large scale ridge and troughs into cut-off pools and their replacement by smaller mobile features in a narrow "zonal" band threading its way between the pools. The accompanying surface changes were complex but usually involved some increase of mobility.

To summarise we may say that the breakdown of the stationary type was associated with a change to a more zonal type of flow, brought about either by cold air being thrust well southwards on the Atlantic or by a degeneration of the large amplitude troughs and ridges.

PART II

Some statistical studies of the deepening and movement of depressions in relation to the thickness pattern in their vicinity.

11. Thermal steering.

11.1. It was remarked in para. 5.1. that though the movement of depressions which swing northwards (Group N) is broadly consistent with the existing steering pattern, the north to south orientation of the thickness lines near the centre is not usually in evidence beforehand but is effected by the circulation of the depression itself. The track charts indicate that depressions of this type swing northwards rather abruptly. In 18 out of the 30 depressions of Class N it was possible to specify definitely the time and position at which this abrupt change in direction occurred: for ease of reference we denote them by H_0 and X respectively. Other times are specified thus: H-24 denotes 24 hours previous to H_0, H_+24 denotes 24 hours after H_0 and so on.

In 16 out of the 18 cases examined the direction of the thickness lines at X backed by 45 degrees or more between H-24 and H_0; the average backing for all cases was approximately 60°.

These cases were segregated into depressions which were on the one hand the first members of a group comprising a spell, and second or subsequent members on the other. It was suggested that the first member should as it were set the thickness pattern for the spell by creating the large-scale trough-ridge pattern over E. Canada, the remaining members of the group being simply steered round the trough. If this were so it would be expected that the backing of the thermal associated with the first member of the group would be greater than with the others. The figures show no significant distinction. A further examination of the charts shows that most spells arise from a series of more or less independent depressions and as each system passes away to northward the eastward or southeastward run of the thickness lines across L is restored, to be distorted again as the next member of the sequence comes on.

/11.2.

11.2. The detailed examination of these 18 cases suggests that the changes in the thermal pattern and in the track of the depression are not quite simultaneous, but that the former of these processes precedes the latter. This suggestion was tested, with the results shown in Fig. 22 which is a frequency table of values of the difference (called the "steering deviation") between the direction of the thickness lines over the centre and the direction of motion of the centre itself. Directions are specified to the nearest two points on the 32-point (compass) scale. Positive values indicate that the depression is moving across the thickness lines from the cold to the warm side, and vice versa. Fig. 22(a) compares the mean direction of movement of the low in the intervals H_{-24} to H_0 and H_0 to H_{+24} with the direction of the thickness lines at (X, H_0). This mean is regarded as the approximate instantaneous direction of movement of the low at (X, H_0). There is evidently some tendency for this movement to be veered in relation to the direction of the thermal over the centre, suggesting that it has not "caught up" with the turning thermal. Fig. 22(b) compares the direction of movement of the low in the interval H_0 to H_{+24} with the thermal at (X, H_0) i.e. over the centre at the beginning of the interval. There is here shown a distinct tendency for the movement of the low to be backed relative to the thermal. This confirms that for the depressions of Type N the backing of the thermal tends to precede the turning of the low - a result which may have some value in short-range forecasting.

Of the four N-class depressions which entered L from the south, three occurred on the east side of a thermal trough of large amplitude and the northward steering of the centre took place in a straightforward way without appreciable distortion of the thickness lines. The fourth system distorted the thermal pattern in the manner discussed above; it was not, however, essentially different from a depression entering L from the west, for it moved up from southwest and crossed the boundary of L very near its southwest corner.

11.3. Histograms of the steering deviation were also prepared for depressions of type C and E. The measurements were made on all possible occasions and not restricted to cases of a marked change of direction. The results are given in Fig. 23 (a) and (b). These each show a distribution curve of approximately normal shape, but slightly skew indicating a tendency in the mean for depression centres to cross the (instantaneous) thickness lines from the cold to the warm side. For the mean movement in the interval H_{-24} to H_{+24} the mode (Fig. 23(a)) is slightly positive, (+2), but for the interval H_0 to H_{+24} (Fig. 23(b)) it is zero. The thermal pattern over the centre at H_0 is a slightly better indicator of the movement of the low in the subsequent 24 hours than of its mean movement in the 48 hours centred on H_0.

11.4. As a test of the usefulness of the principle of thermal steering we evaluate the percentages of steering deviations having the value **zero** and lying within the range $\pm$ 2 on the scale used i.e. less than 11° and 34° respectively. These are as follows:-

Steering deviation	Mean movement of low from H_{-24} to H_{+24}	Movement of low from H_0 to H_{+24}
$\leqslant 11^\circ$	12%	25%
$\leqslant 34^\circ$	48%	57%

Thus the crude application of the thermal steering principle gives a satisfactory guide to the movement of the depression in only about half the total number of cases.

/12.

12. The movement of depressions in relation to their deepening and filling.

It has been suggested that there may be a very broad connexion between the rate of deepening of a depression and its speed of movement, rapidly deepening systems tending to be slow-moving while fast moving systems show little tendency to deepen rapidly. To test this, certain data were extracted for major systems for those periods in which they were uncomplicated by processes of splitting or amalgamating with other systems including secondaries and breakaways, that is, only simple systems, indentifiable over a time interval of at least 12 hours, were considered. The data extracted were:-

(i) The movement of the centre in hundreds of miles over a 12-hour period.

(ii) The change of central pressure over the same period.

This was done for two separate periods of three months in the winter half year, viz. January to March and October to December, 1950. The results are shown in the double entry frequency tables, Tables III (a) and (b). The two tables are almost exactly comparable, the total numbers of cases being respectively 191 and 186.

Table III. Frequency table showing change of pressure in relation to the movement of depressions over 12-hour periods.

(a) January - March, 1950

Movement of centre in 12 hours (100's of miles)	Change of pressure in 12 hours (mb.)									
	+13	+9 to +12	+5 to +8	+1 to +4	0	-1 to -4	-5 to -8	-9 to -12	-13	Total
0 - 1	0	0	2	3	1	0	0	1	1	8
2 - 3	5	1	5	17	15	21	14	3	2	83
4 - 5	1	1	3	5	15	16	6	7	6	60
6 - 7	0	1	2	4	7	7	8	5	0	34
8 - 9	0	1	0	0	1	1	0	0	0	3
10	0	0	1	0	0	0	1	1	0	3
Mean movement in 12 hours (100's of miles)	2.8	5.5	3.9	3.2	4.1	4.0	4.2	5.1	3.6	191

/(b)

(b) October to December, 1950

Movement of centre in 12 hours (100's of miles)	Change of pressure in 12 hours (mb.)									
	13	+9 to +12	+5 to +8	+1 to +4	0	-1 to -4	-5 to -8	-9 to -12	-13	Total
0 - 1	0	0	2	2	6	4	1	0	1	16
2 - 3	2	2	9	27	21	21	14	5	1	102
4 - 5	1	0	3	9	16	12	4	3	4	52
6 - 7	0	0	0	0	6	4	1	1	0	12
8 - 9	0	0	0	0	1	0	2	0	0	3
10	0	0	0	0	1	0	0	0	0	1
Mean movement in 12 hours (100's of miles)	3.2	2.5	2.6	2.9	3.6	3.3	3.5	3.6	3.5	186

In the bottom row of each table is entered the mean speed of movement for each rate of deepening. No relation is apparent apart from a slight tendency for deepening depressions to move a little faster than filling ones. There is no tendency for rapidly deepening systems to move slowly. On the other hand it is well known that very deep (occluded) systems are often slow moving: these results suggest that the slowing down often occurs after the process of deepening.

The last columns of Table IV (a) and (b) shows that about half the movements considered were at speeds of about 15 to 30 knots. Very small movements were rare.

13. The deepening of depressions in relation to certain thermal features.

13.1.(a) Thermal gradient over the centre.

During the period January to March, 1950 among the systems studied there were 25 cases of rapid deepening (10 mb. or over during a 12-hour interval). The thickness gradient over the centre at the beginning of the 12-hour interval was measured and compared with the five-year mean thickness gradient at the same place for the appropriate month. The result is shown in Fig. 24 which is a frequency curve of values of the ratio actual thickness gradient to this "normal" gradient. The values of this ratio varied from 1.0 to 2.2 with a mean of 1.7. There was no indication that the greater rates of deepening were associated with higher (or lower) values of this ratio.

Though the number of data is limited it is probably significant that in 4/5 of the total number of cases the ratio was equal to or greater than 1.5.

13.2. Thickness anomaly over the centre.

The thickness anomaly (actual thickness minus five-year mean) over the deepening centre was evaluated for the 25 cases in 13.1. Fig. 25 shows the resulting frequency curve. The thickness anomalies are expressed in feet for ranges centred on the values 0, ± 100, ± 200 feet etc. (the ranges are -50 to +50, ± 60 to ± 150 and so on). The values show a slight bias towards positive values of the anomaly (14 cases of positive anomaly, 6 of negative). The mean value is + 120 feet.

/It

It might have been expected that, since vertical instability is to some extent favourable for the deepening of depressions (Sumner 1950), a greater tendency towards negative thickness anomalies would have been apparent in these cases.

13.3. Relation of deepening depressions to thermal patterns models.

The 25 cases of rapid deepening in 13.1 above were examined qualitatively in relation to the thermal pattern in their vicinity. In all cases the thermal patterns were qualitatively consistent with the developmental ideas put forward by Sutcliffe (1947, 1950) in the sense that the deepening was located in the "correct" relation to thermal troughs and ridges, difluence and confluence. But the degree of deepening appeared to bear no easily recognisable relationship to the intensity of development estimated by eye from the thermal pattern, nor was there any easily recognisable pointer to the exact location of the deepening. It is evident that for further increase in accuracy of estimating development quantitative treatment of the thickness pattern is at least necessary, though it may not be sufficient.

14. Conclusions.

The following conclusions are drawn from this study.

(a) Depressions moving through the selected area (L) exhibit many of the features of evolution common to middle latitude systems. While the systems move on the whole more or less eastward there is an important class which turns away sharply northward or northwestward and become slow moving. It is suggested that the peculiarity of the area in this respect arises from the presence of the Greenland massif which causes a sharper separation between the two classes than in localities where topographical factors are absent or less pronounced..

(b) The behaviour of depressions is not random: when the sytems are classified according to their movement the various classes seem to occur in rather loosely defined spells or groups lasting from a few days to a week or more.

(c) Some of these spells are associated with characteristic hemispherical upper flow patterns. The most definite of these is the pattern of low moving large amplitude troughs and ridges associated with the stagnating depressions referred to in (a). Such a pattern usually persists throughout each spell, with persistent cyclonic or anticyclonic conditions in certain areas, but the geographical locations of the upper troughs and ridges and the corresponding surface systems vary from one spell to another.

(d) The greatest degree of steadiness in these well defined upper flow patterns is often not attained until towards the end of the spell - that is after the surface patterns have themselves become steady. This suggests that the upper flow pattern is not the antecedent in any causal relation there may be between it and the surface patterns.

(e) Regarded qualitatively the large scale upper flow patterns afford no consistent indications of impending changes of type such as would be useful in medium range forecasting.

(f) The evolution of the surface systems usually conforms to the thermal developmental models put forward by Sutcliffe (1947, 1950), but in most cases the thermal and surface patterns evolve together and it is not possible to forecast one from the other.

/(g)

(g) Differences in the behaviour of individual depressions appear to depend upon rather subtle factors including perhaps details of the thermal structure which are not apparent in a broad survey. It is an open question whether the quantitative study of thickness patterns would be a useful aid to forecasting in the area studied.

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Sutcliffe, R.C. and Forsdyke, A.G.	1950	Quart. J.R. Met. Soc. London 76, p. 189.

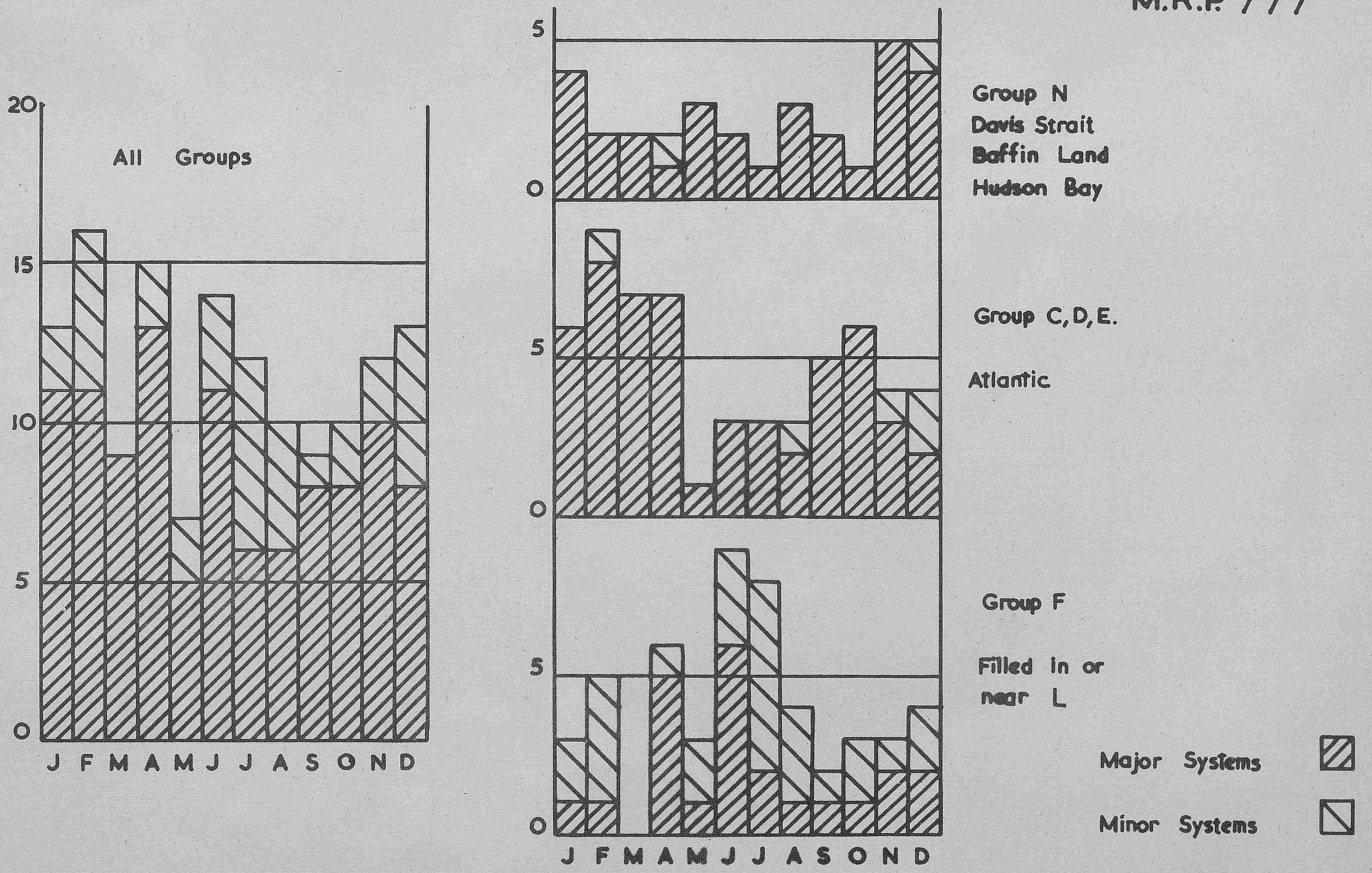
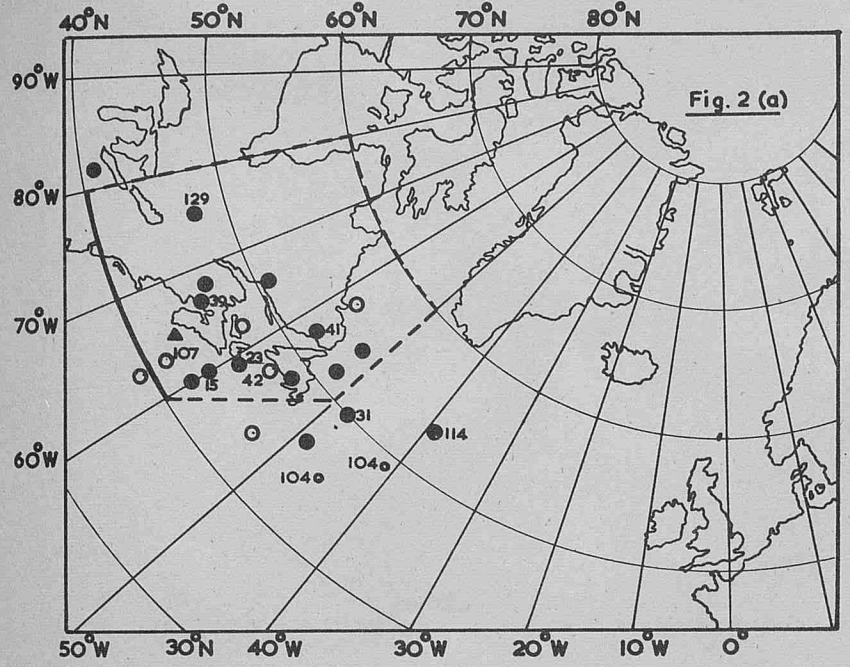
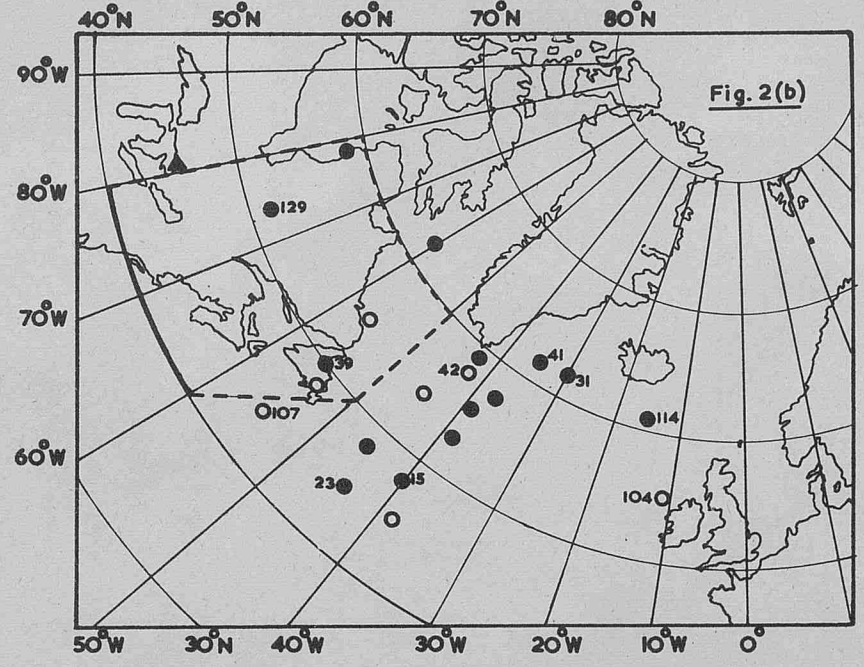
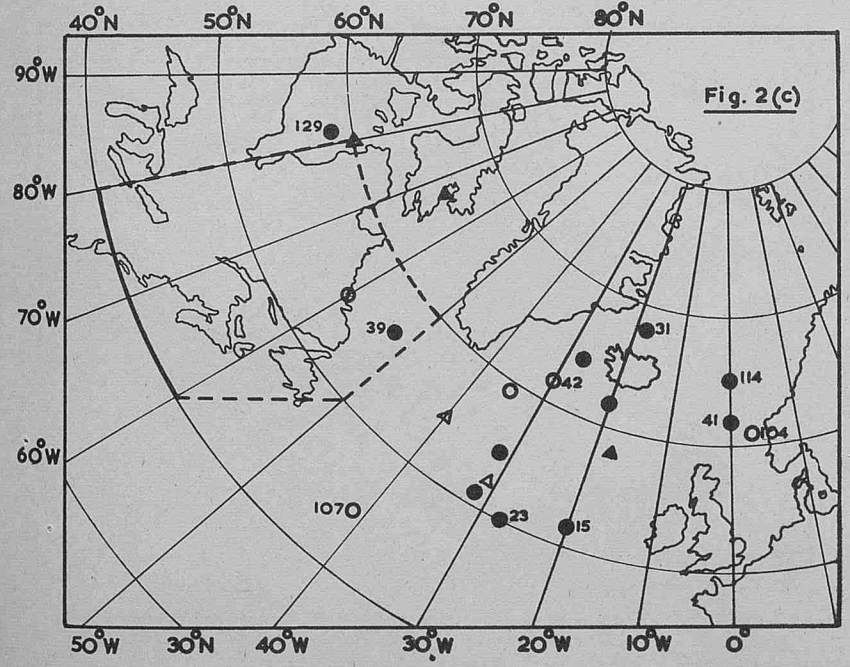


Fig. 1. Monthly frequency of depressions traversing area L (1950)







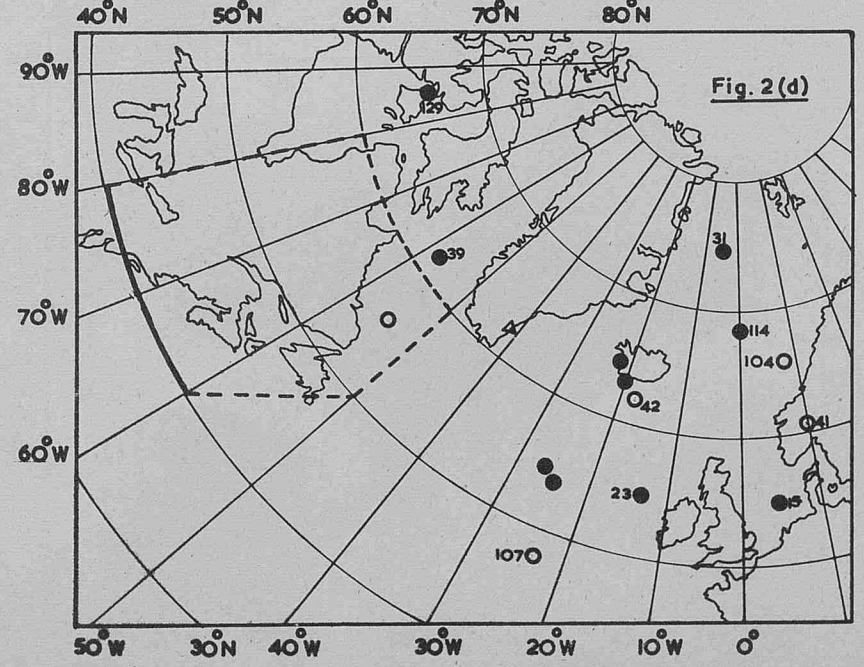


Fig. 2

Positions of centres of depressions which entered area L from South:–

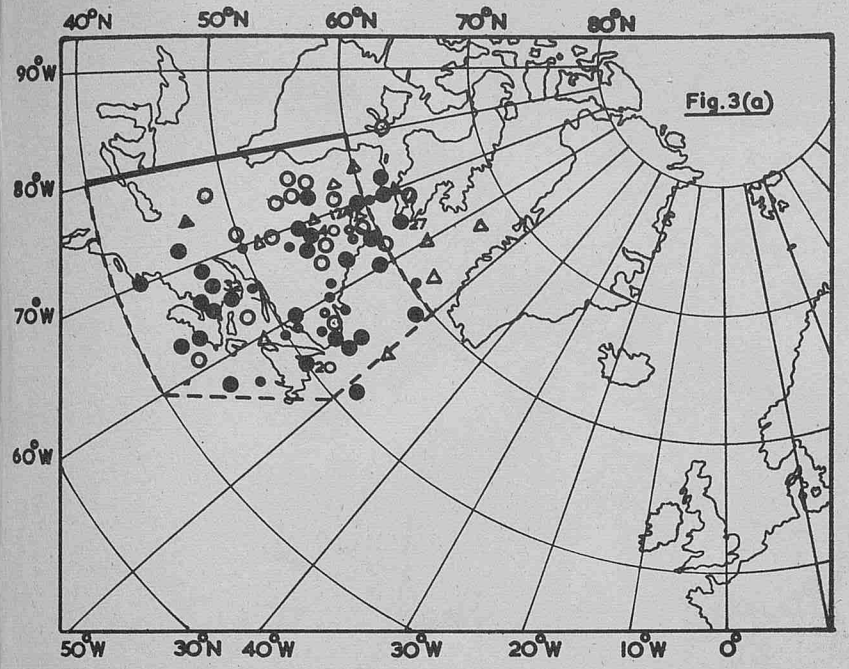
Fig. 2(a)	1 day	after entering L
Fig. 2(b)	2 days	
Fig. 2(c)	3 days	
Fig. 2(d)	4 days	

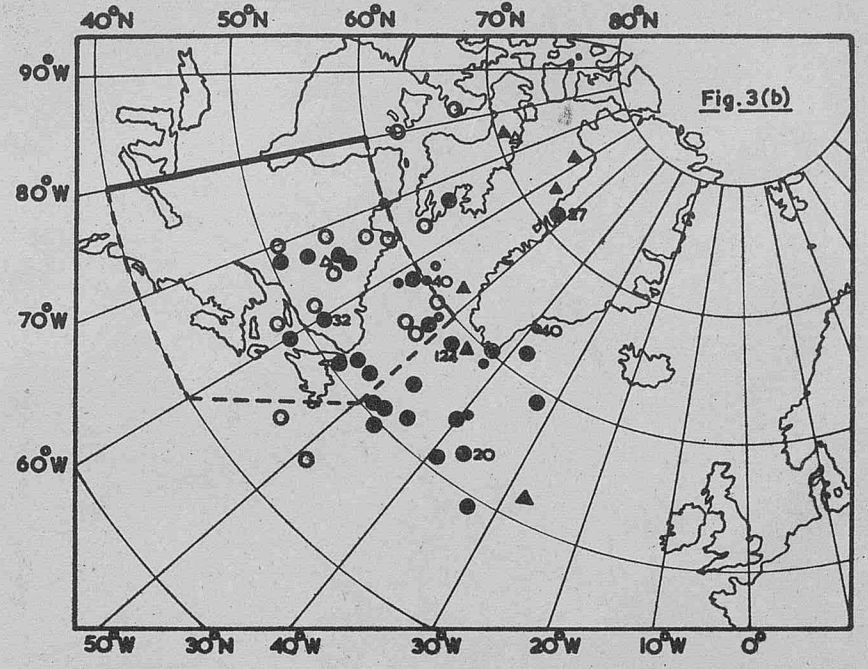
● Winter (Oct. to March).

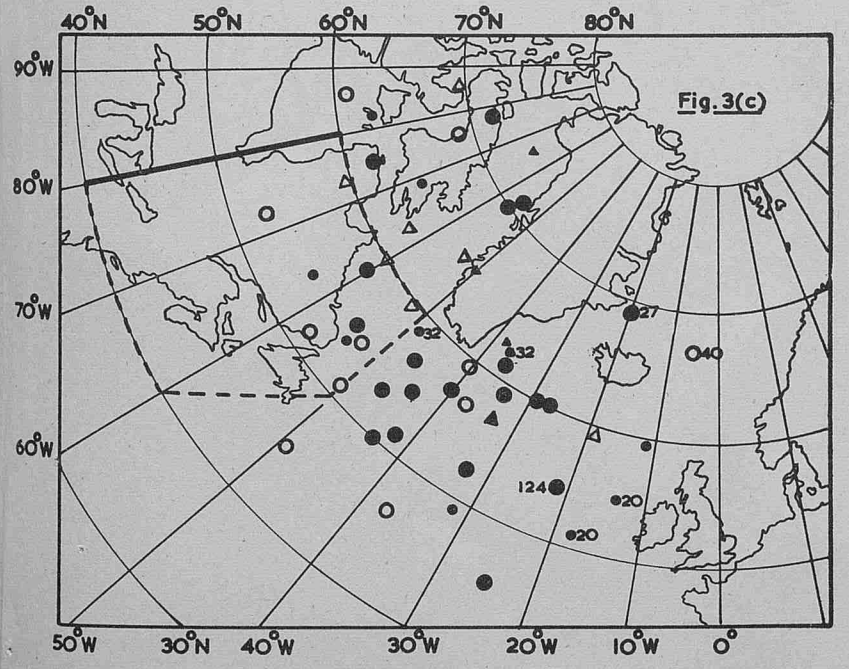
O Summer (Apr. to Sept.).

▲, △ denote system filled

•, o denote centres of system which divided after entering L







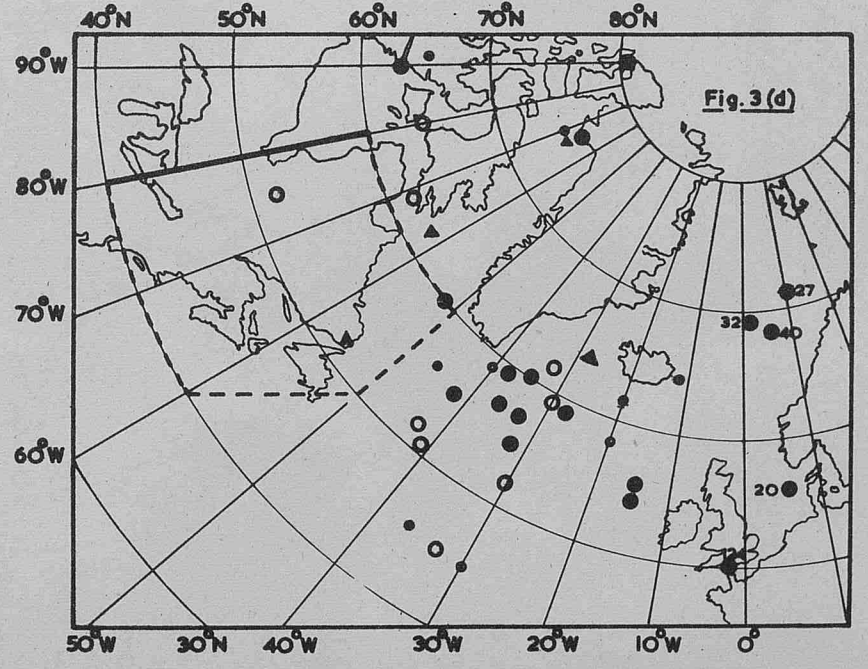


Fig. 3

Positions of centres of depressions which entered area L from West:-

Fig. 3(a)	1 day	after entering L
Fig. 3(b)	2 days	
Fig. 3(c)	3 days	
Fig. 3(d)	4 days	

● Winter (Oct. to March).

○ Summer (Apr. to Sept.).

▲, △ denote system filled.

•, ○ denote centres of system which divided after entering L

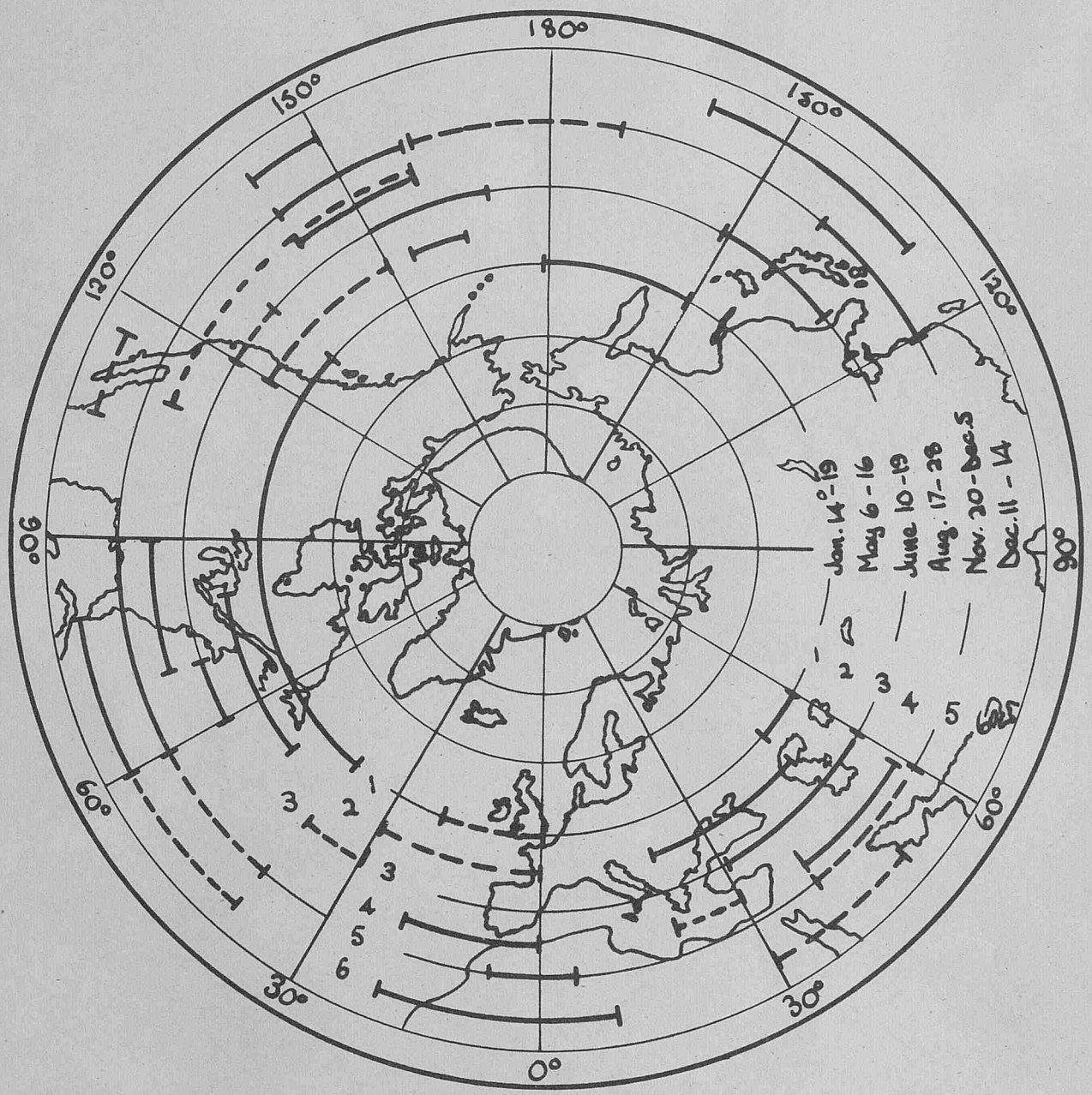


Fig. 4. Sectors occupied by persistent upper troughs and ridges in N I spells.

Troughs ——— Ridges - - - -

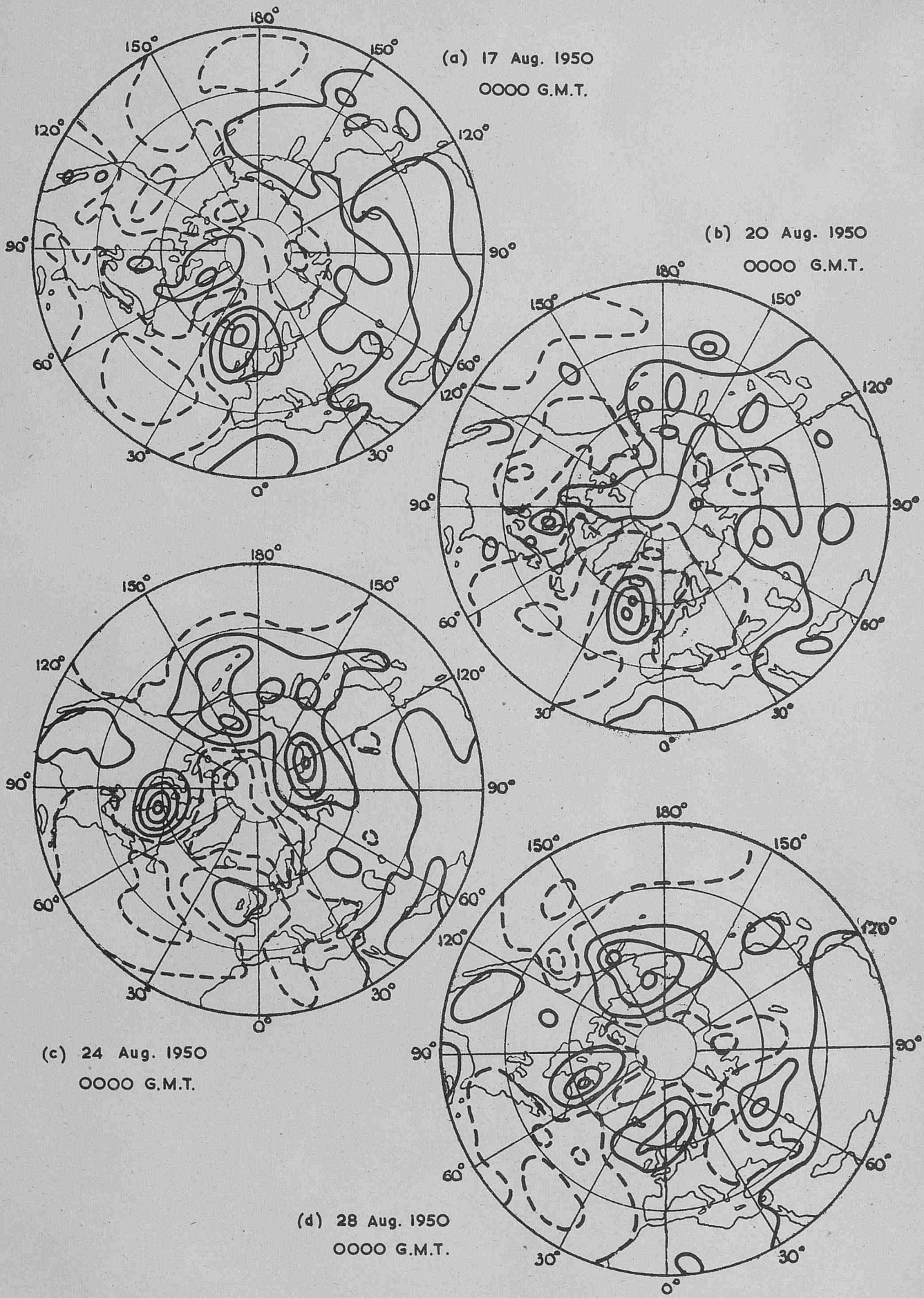


Fig.5. Surface charts for NI spell 17–28 Aug. 1950.

Isobars at 8mb. intervals.
1016 mb. and above — — —
1008 mb. and below ————

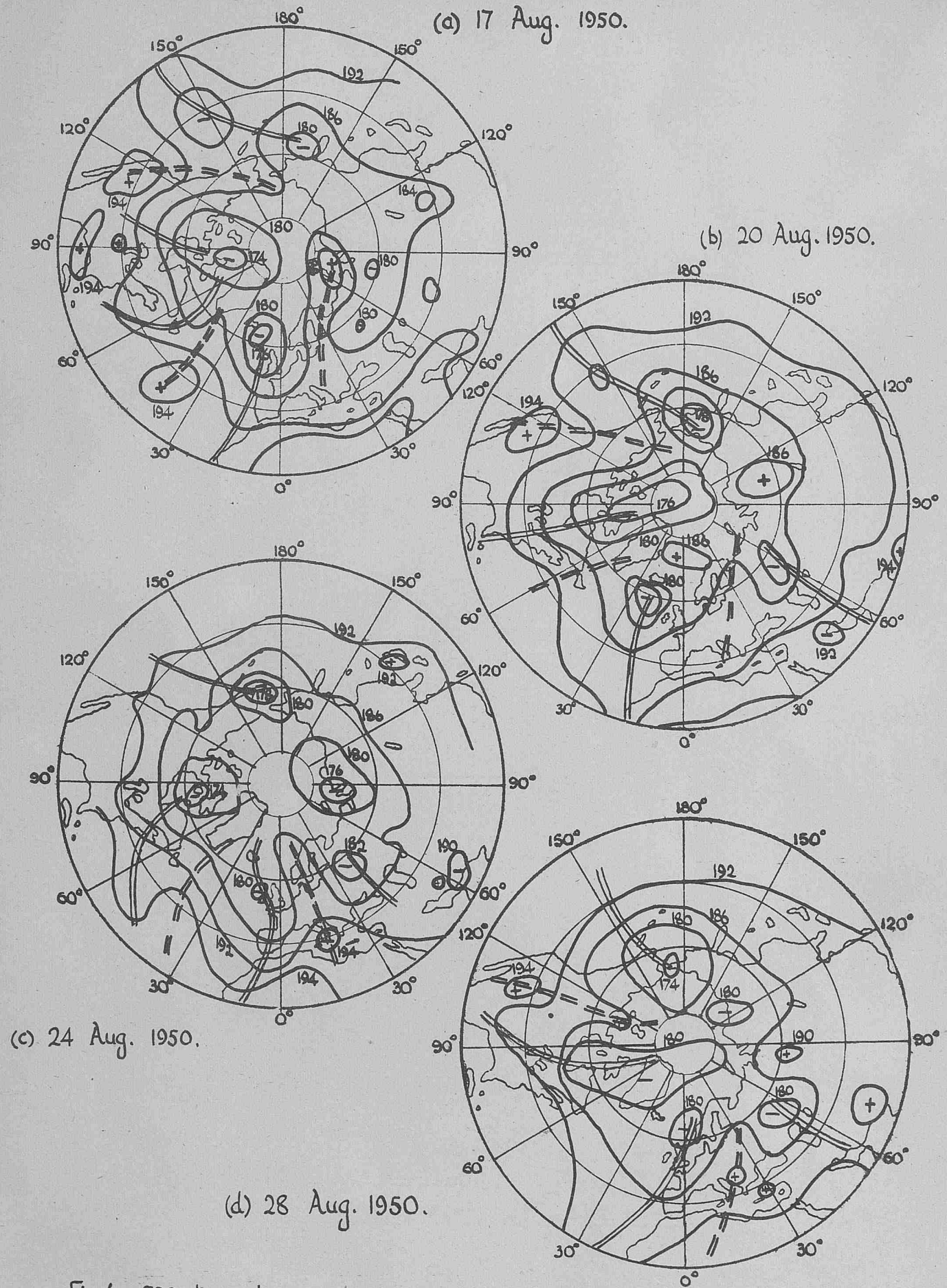


Fig. 6. 500 mb. contours at 0300 G.M.T. for N1 spell 17 – 28 Aug. 1950
Contour heights in hundreds of feet.
Persistent trough lines ———
Persistant ridge lines =====

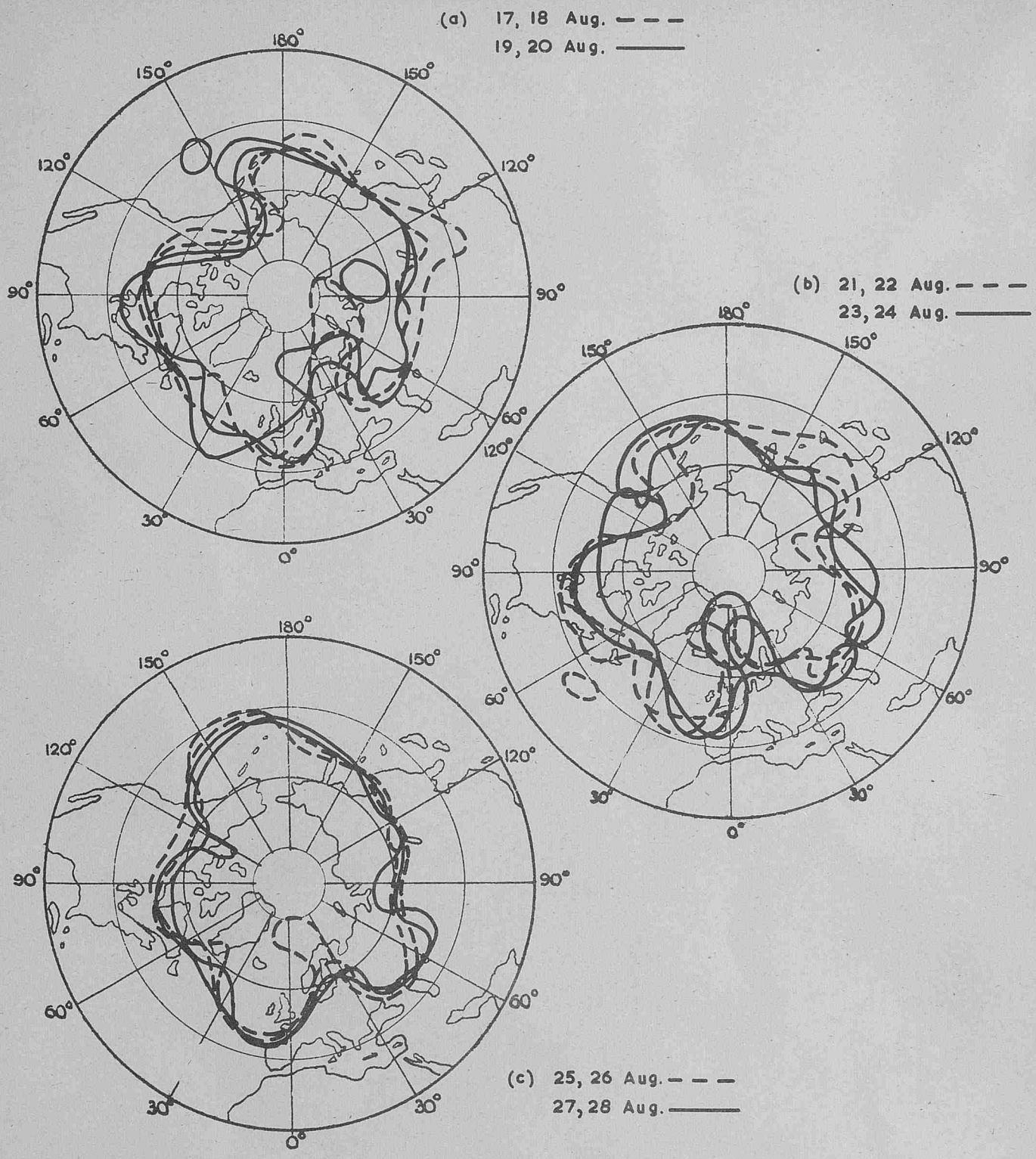


Fig. 7. Nispell 17–28 Aug. 1950. Composite charts of 18600 ft. line in 500 mb. pattern at 0300 G.M.T.

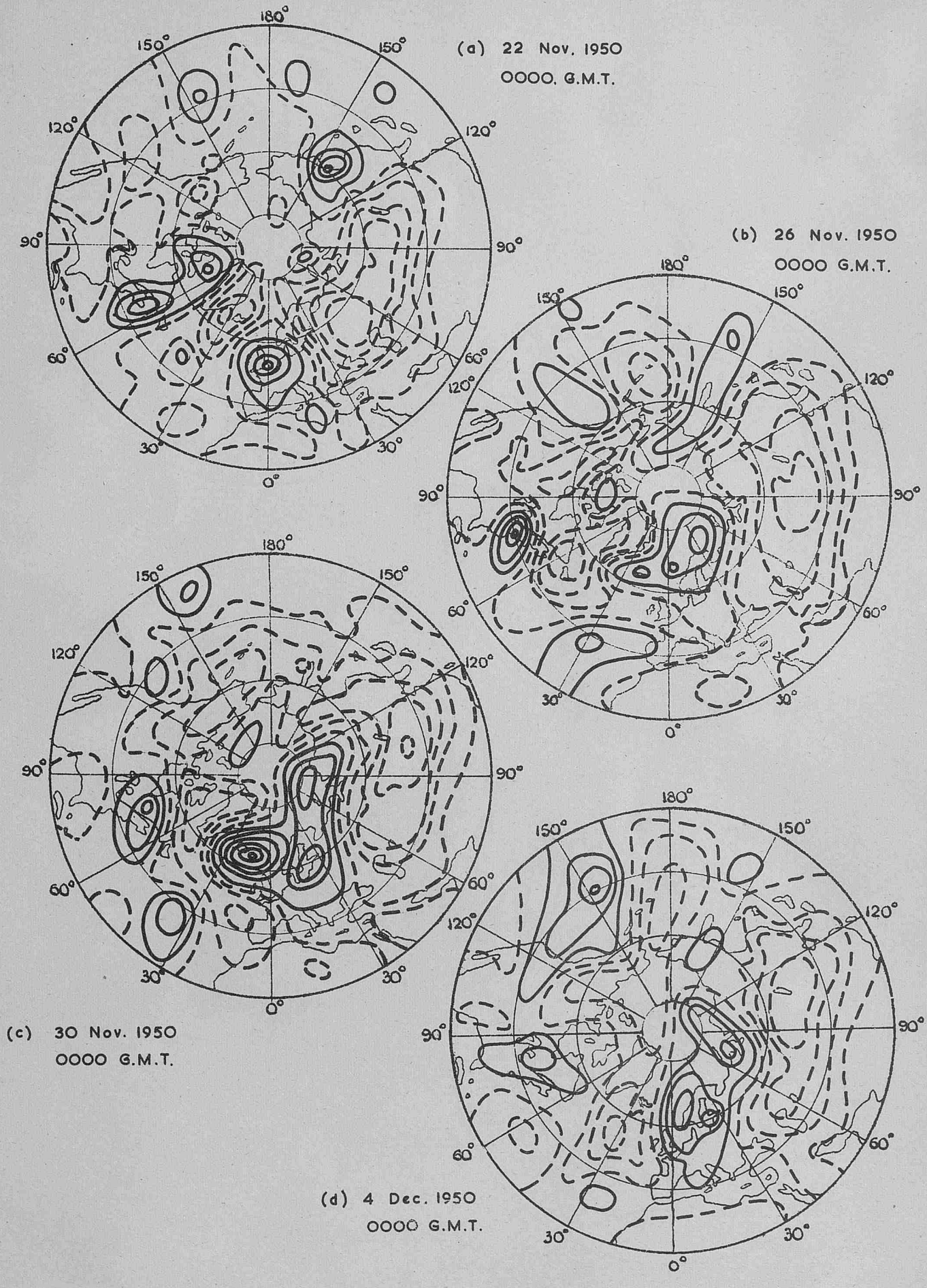


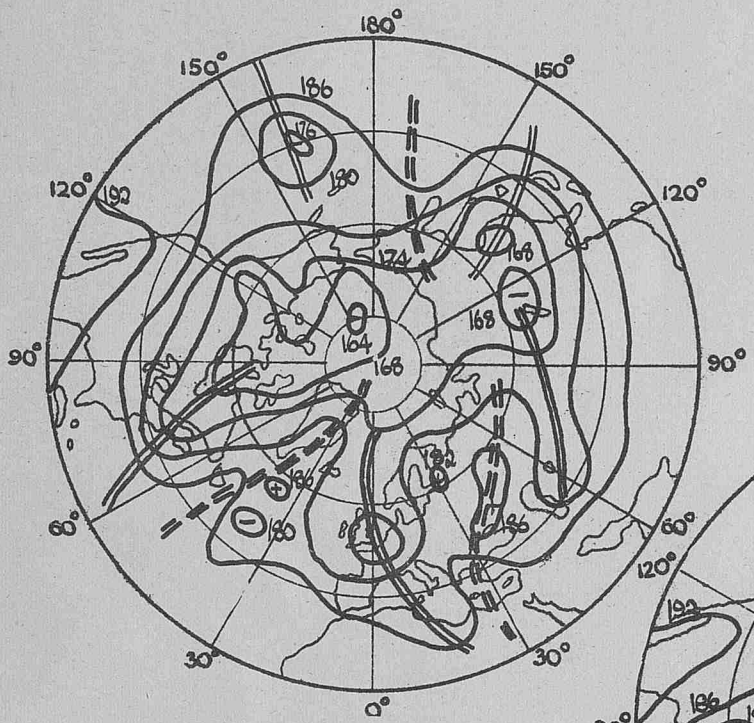
Fig. 8. Surface charts for NI spell 20 Nov.– 5 Dec. 1950

Isobars at 8mb. intervals

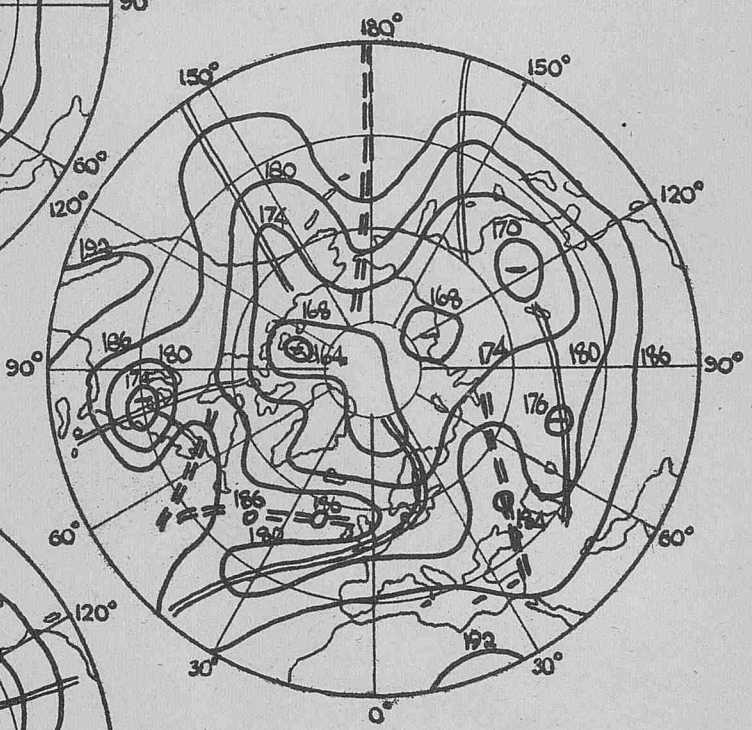
1016mb. and above — — —

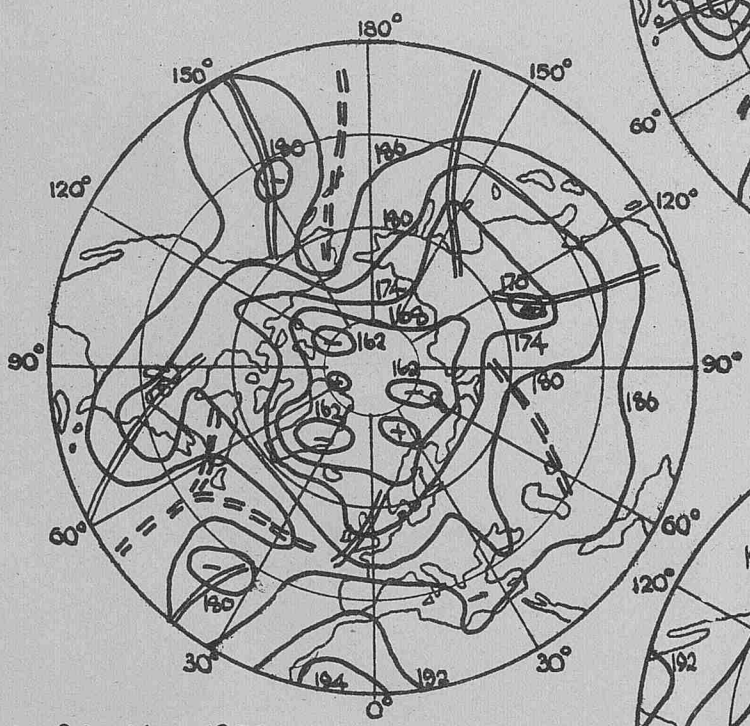
1008mb. and below ———

(a) 22 Nov. 1950.

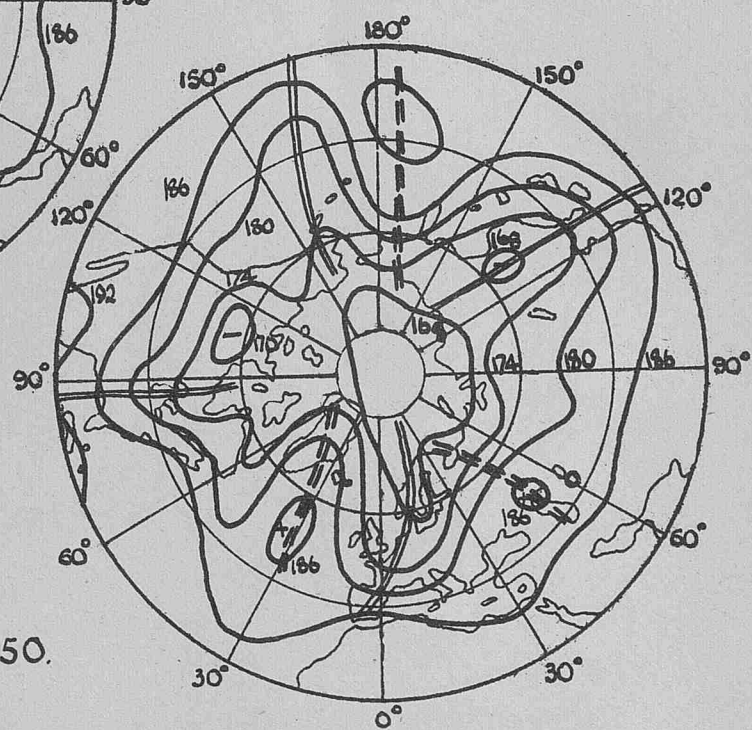


(b) 26 Nov. 1950.





(c) 30 Nov. 1950



(d) 4 Dec. 1950.

Fig. 9. 500 mb contours at 0300 G.M.T. for N1 spell, 20 Nov. – 5 Dec 1950.
Contour heights in hundreds of feet
Persistent trough lines ——— Persistent ridge lines. = = = = =

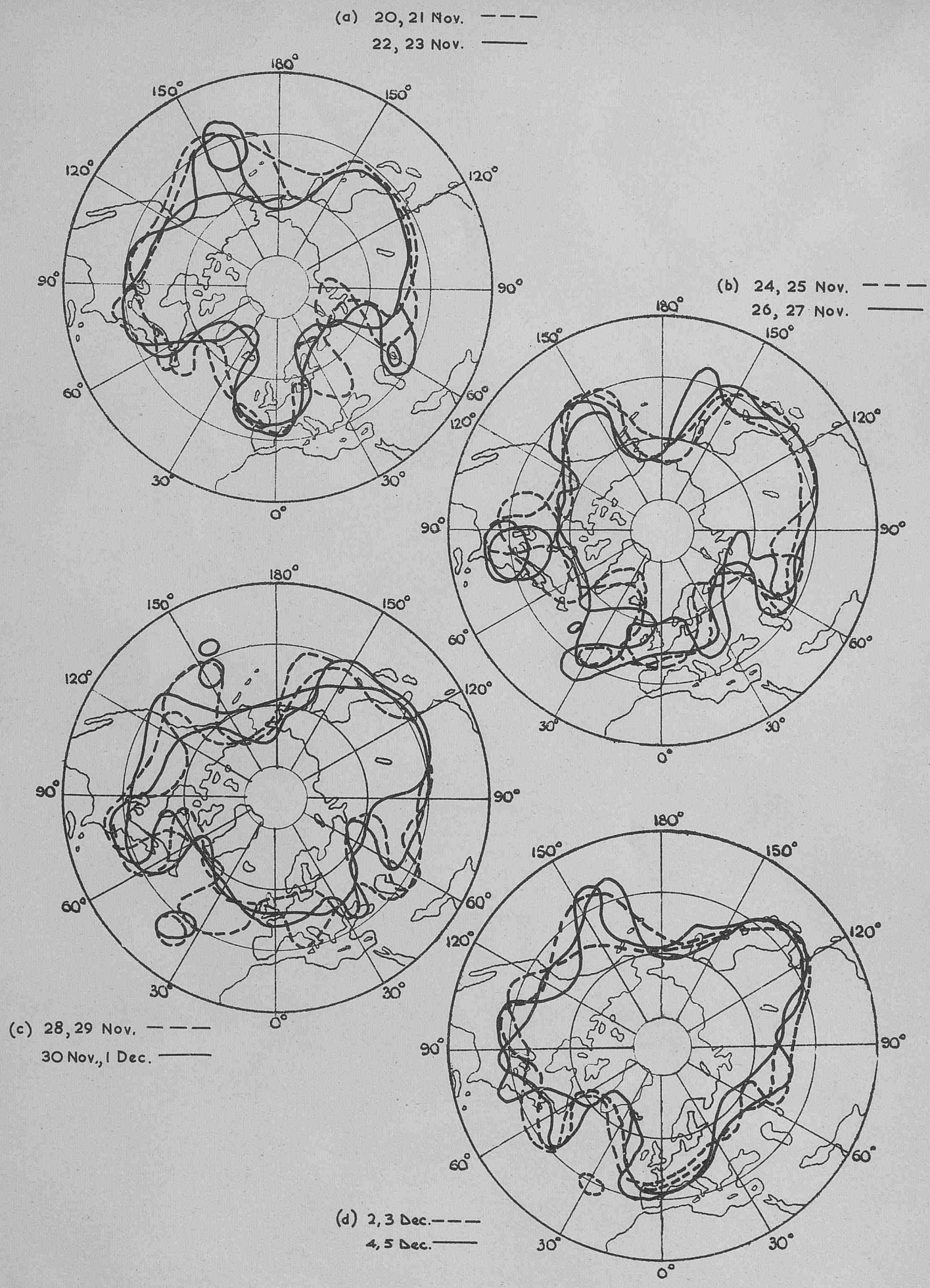


Fig. 10. N1 spell, 20 Nov. – 5 Dec. 1950. Composite charts of 18,600 ft. line in 500 mb. contours for 0300 G.M.T.

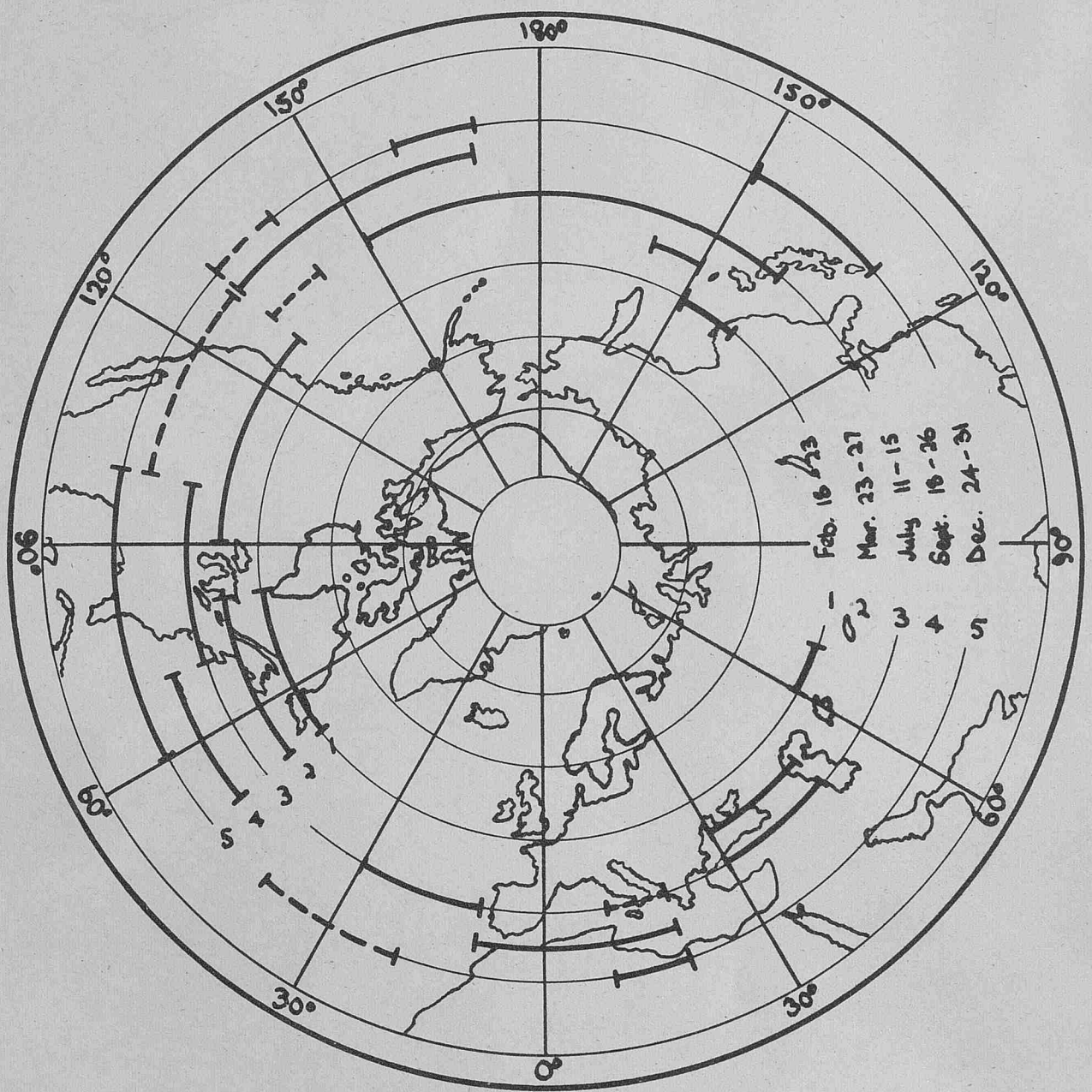


Fig. 11. Sectors occupied by persistent upper troughs and ridges in N 2 spells, 1950.

Troughs ——— Ridges - - - -

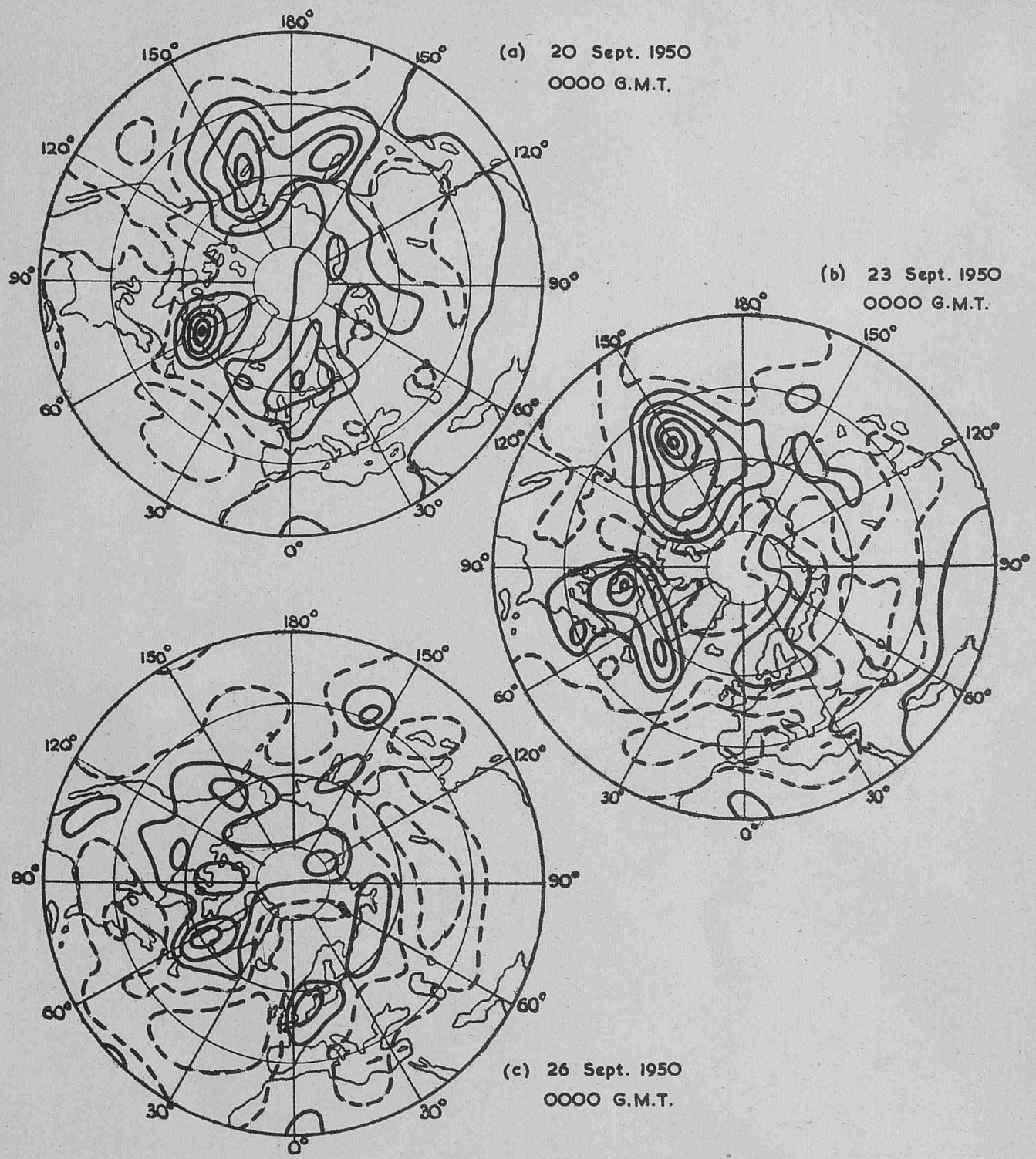


Fig. 12 Surface charts for N2 spell 18–26 Sept. 1950

Isobars at 8 mb. intervals.

1016 mb. and above — — —

1008 mb. and below ————

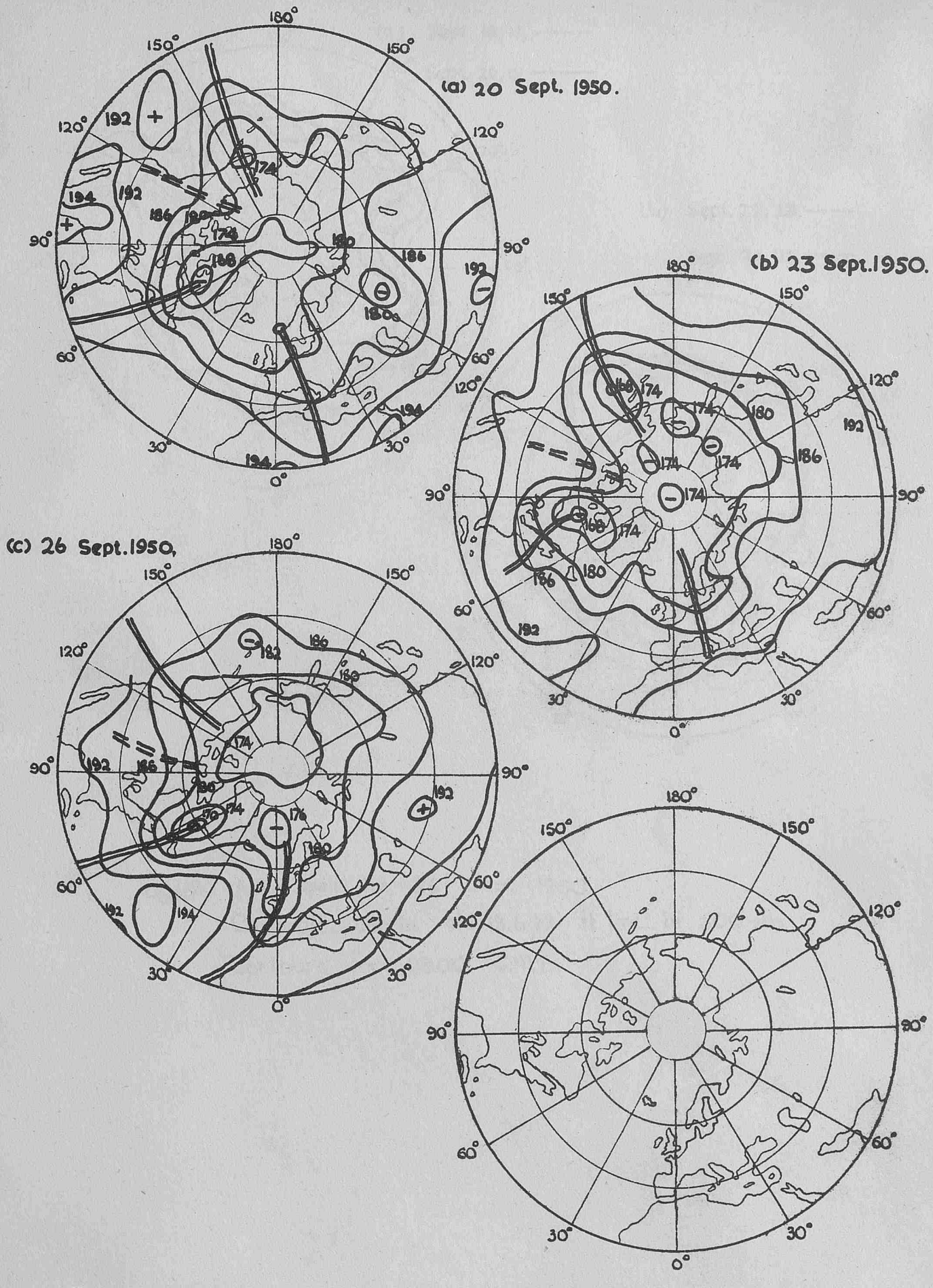


Fig.13. 500 mb Contours at 0300 G.M.T. for N2 spell 18-26 Sept. 1950
Contour heights in hundreds of feet.
Persistent trough lines ═══ Persistent ridge lines ======

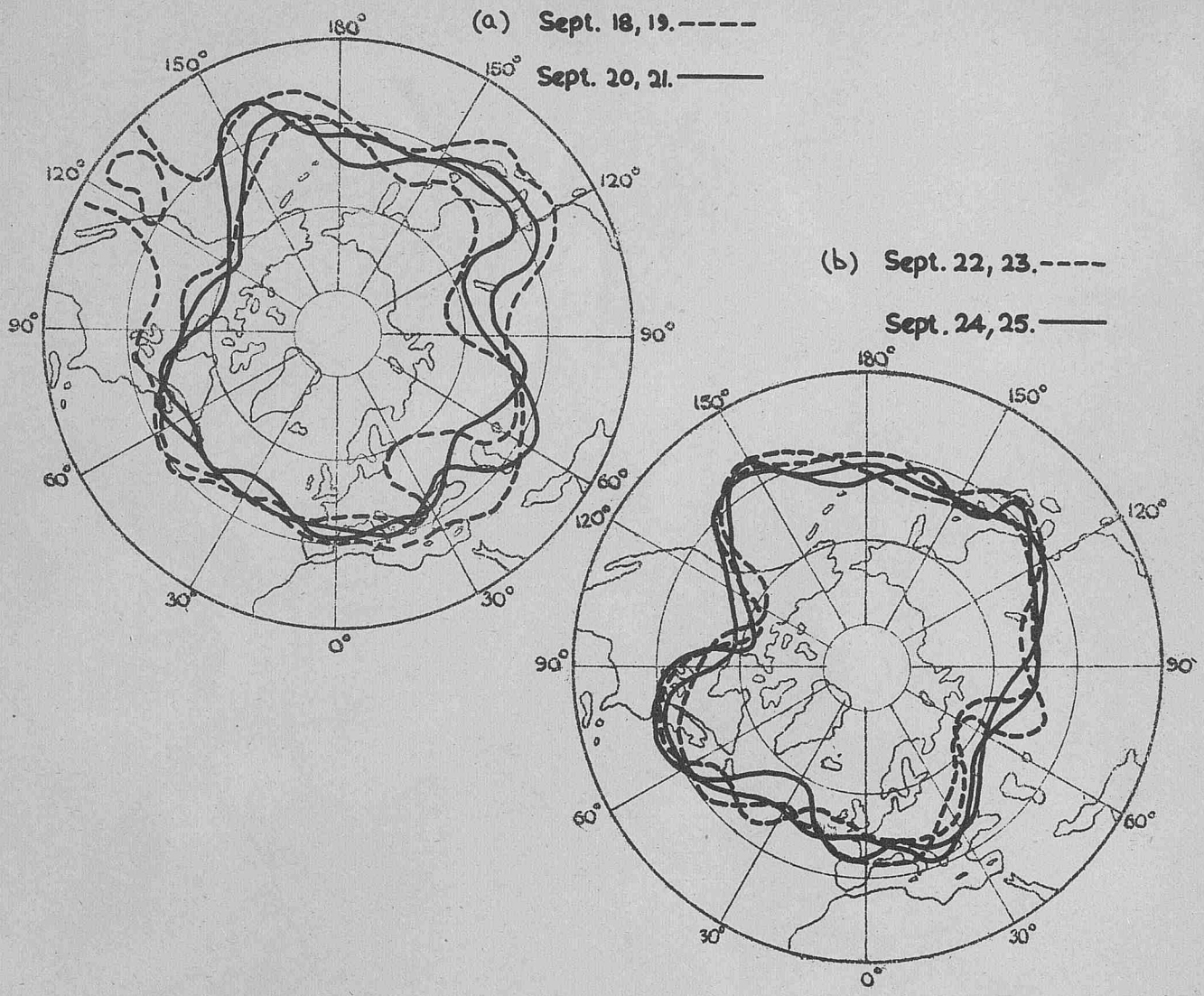


Fig. 14 N2 spell 18 – 26 Sept. 1950
Composite charts of 18,600 ft. line in 500 mb. contours for 0300 G.M.T.

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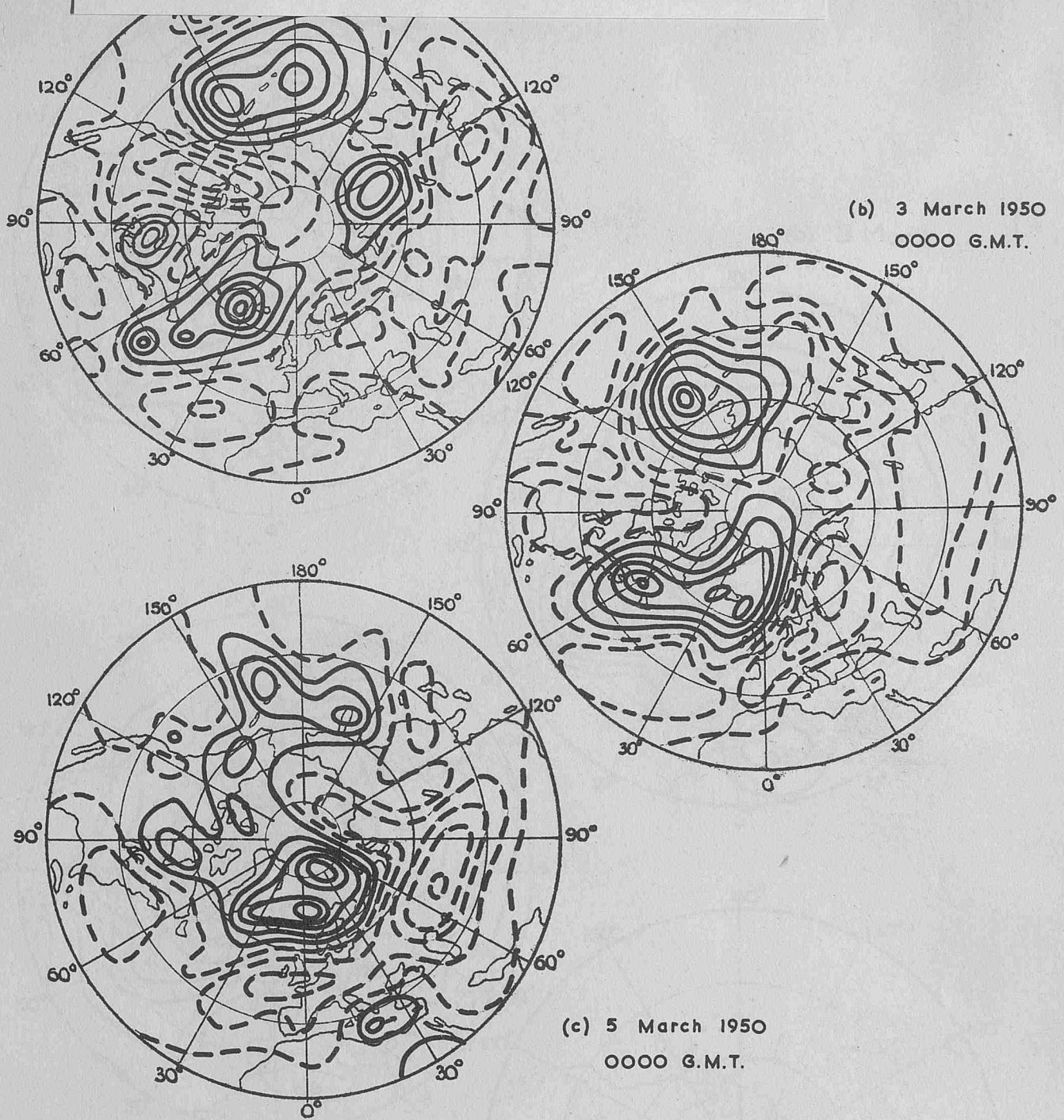


Fig.15 Surface charts for C spell 24 Feb.– 4 March 1950

Isobars at 8mb. intervals

1016mb. and above — — —

1008mb. and below ————

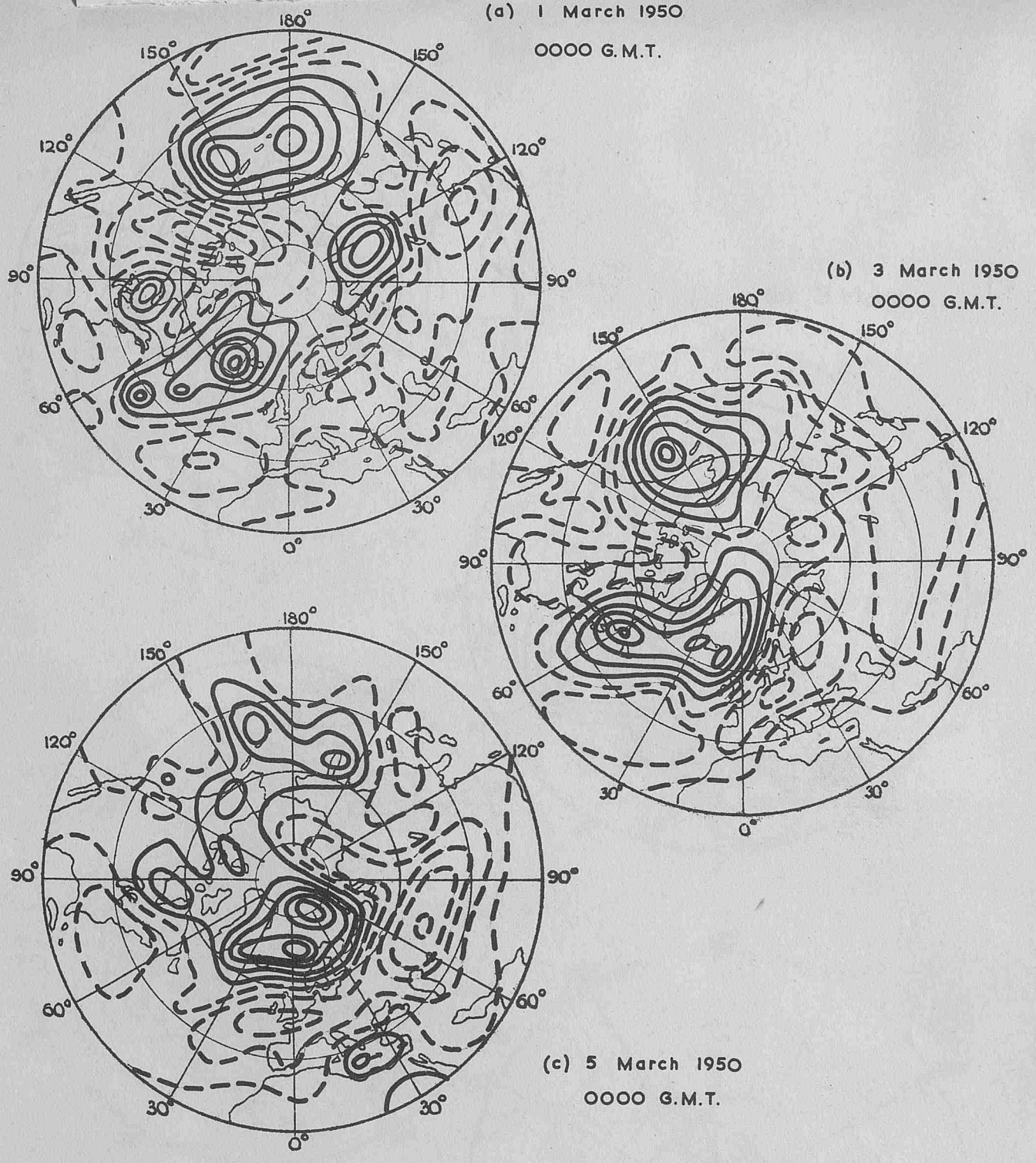


Fig.15 Surface charts for C spell 24 Feb.– 4 March 1950

Isobars at 8mb. intervals

1016mb. and above — — —

1008mb. and below ————

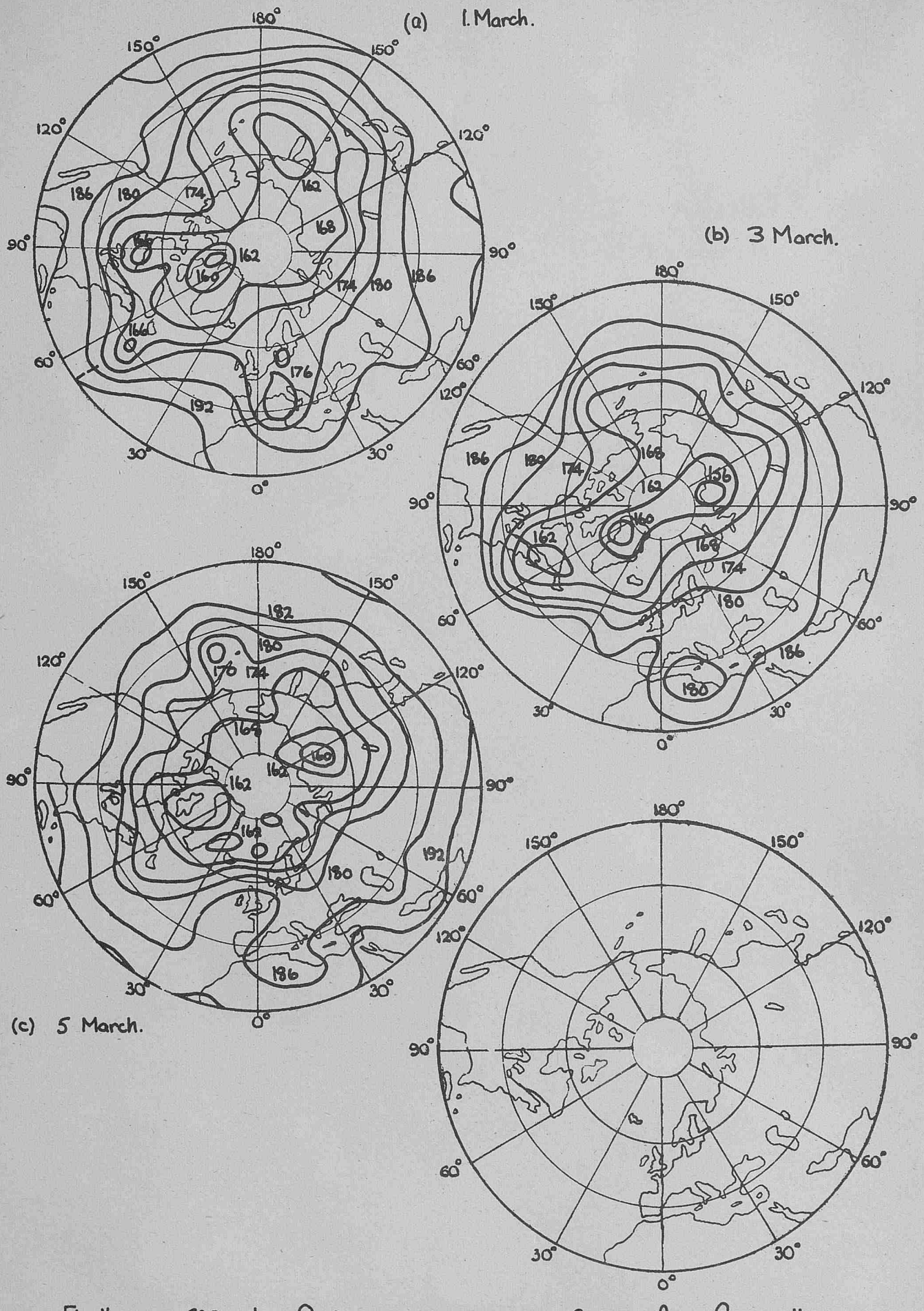


Fig. 16. 500 mb. Contours at 0300 G.M.T. for C spell
24 Feb. – 4 March 1950
Contour heights in hundreds of feet.

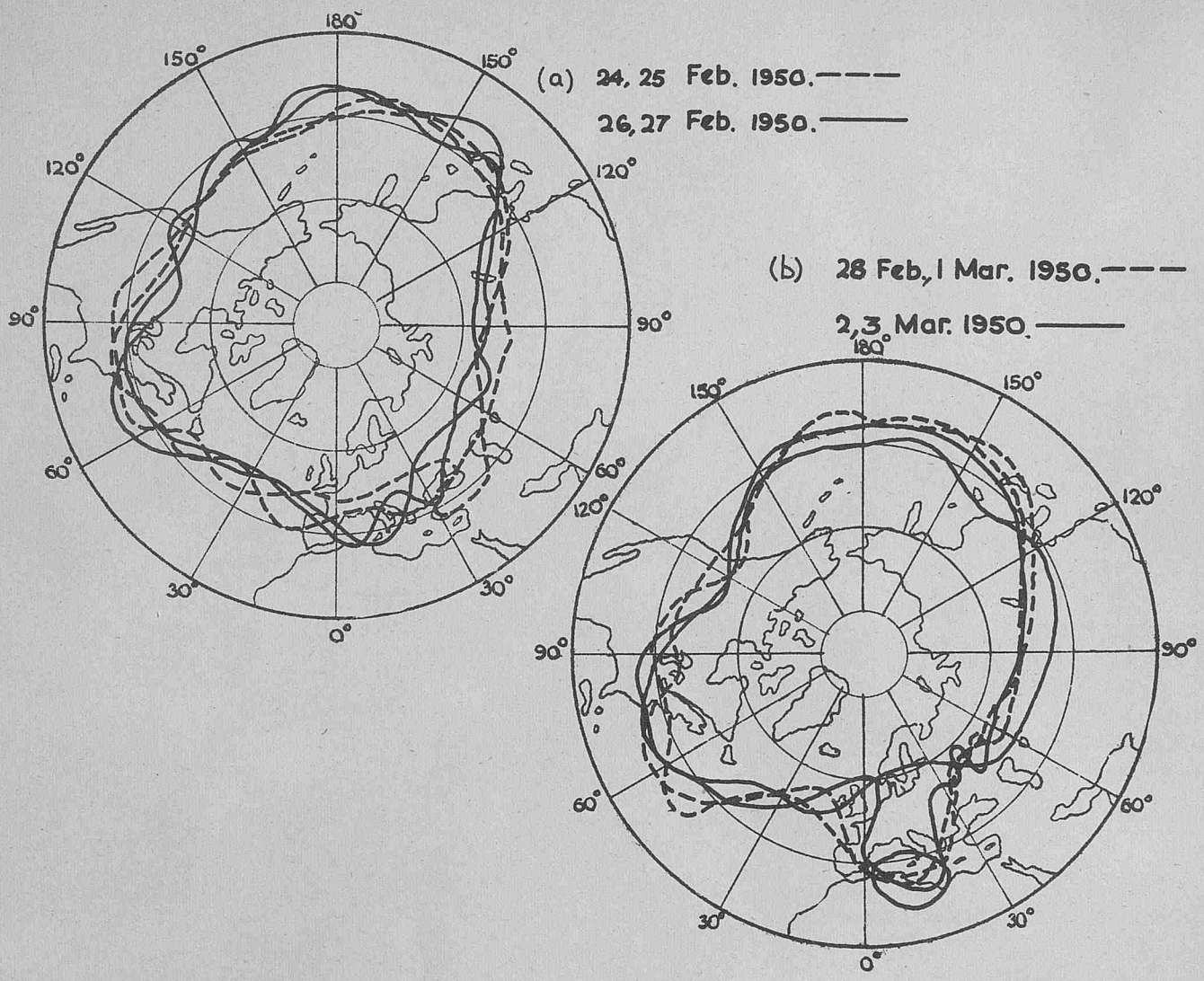


Fig.17. C spell, 24 Feb.– 4 March 1950.

Composite charts of 18,000 ft. line in 500 mb. contours for 0300 G.M.T.

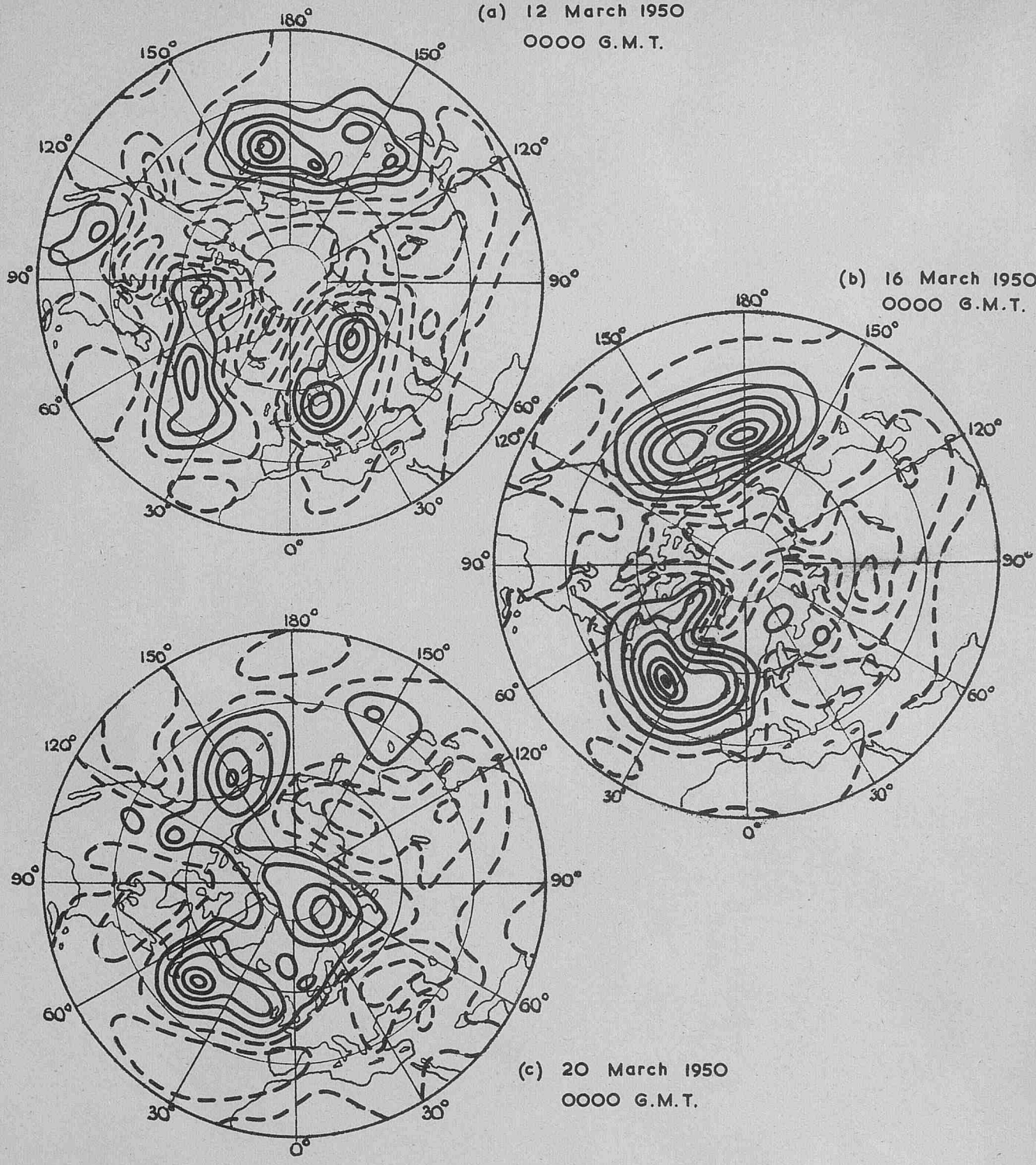


Fig. 18 Surface charts for E spell 12–20 March 1950

Isobars at 8 mb. intervals

1016 mb. and above — — —

1008 mb. and below ————

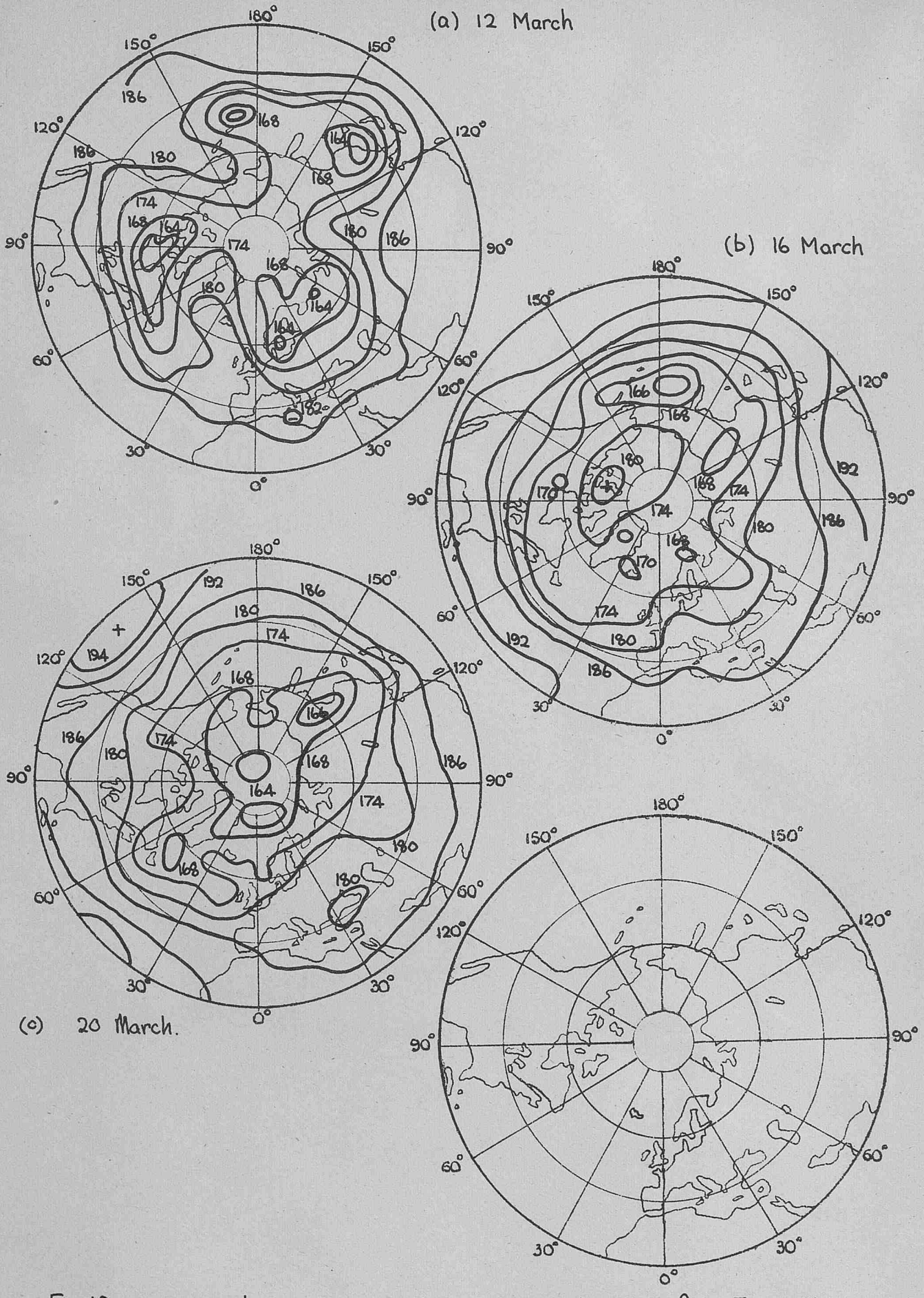


Fig. 19. 500 mb. contours at 0300 G.M.T. for E spell. 12 - 20 March 1950. Contour heights in hundreds of feet.

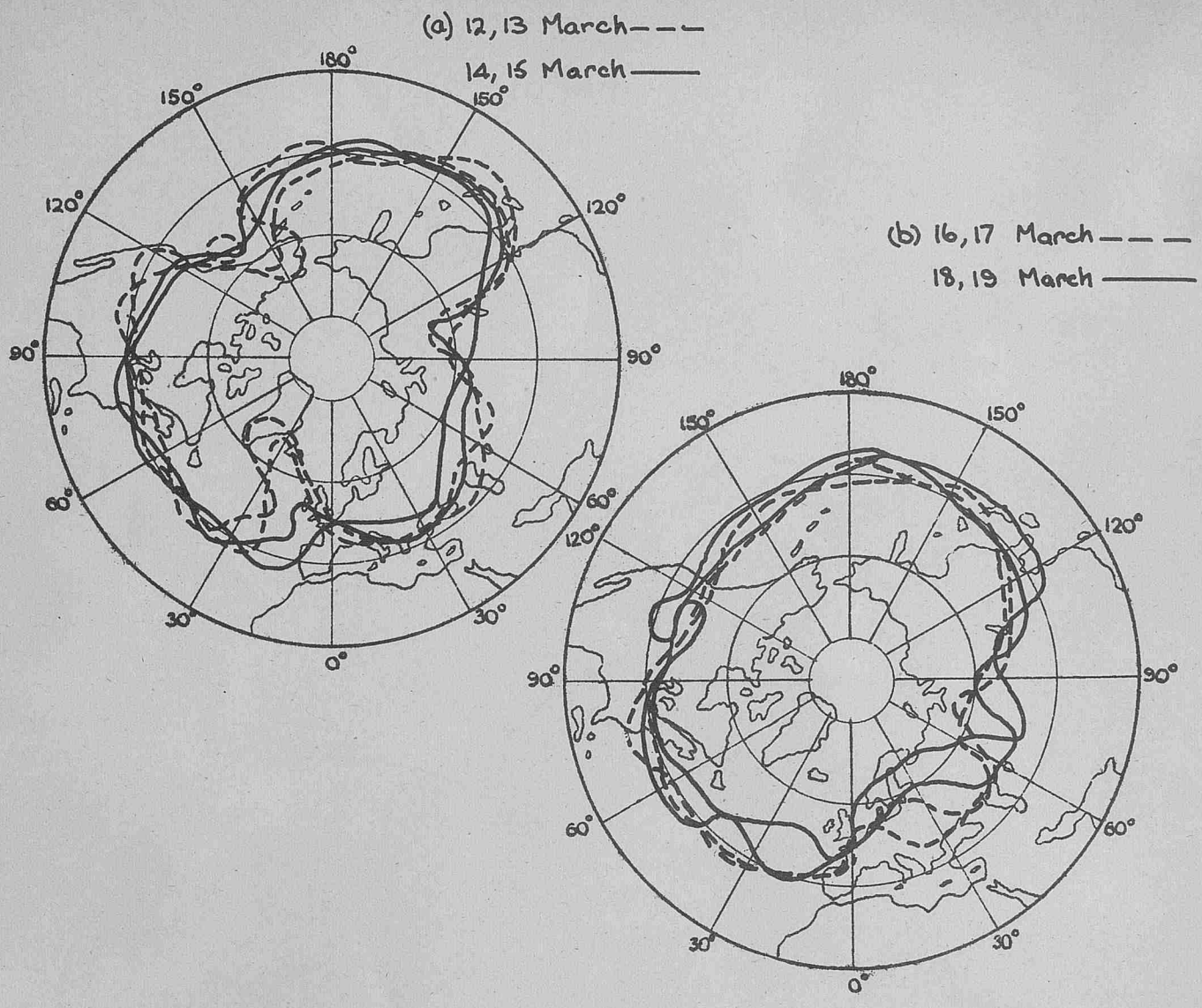


Fig. 20. E spell, 12 – 20 March 1950.

Composite charts of 18,000 ft. line in 500 mb. contours for 0300 G.M.T.

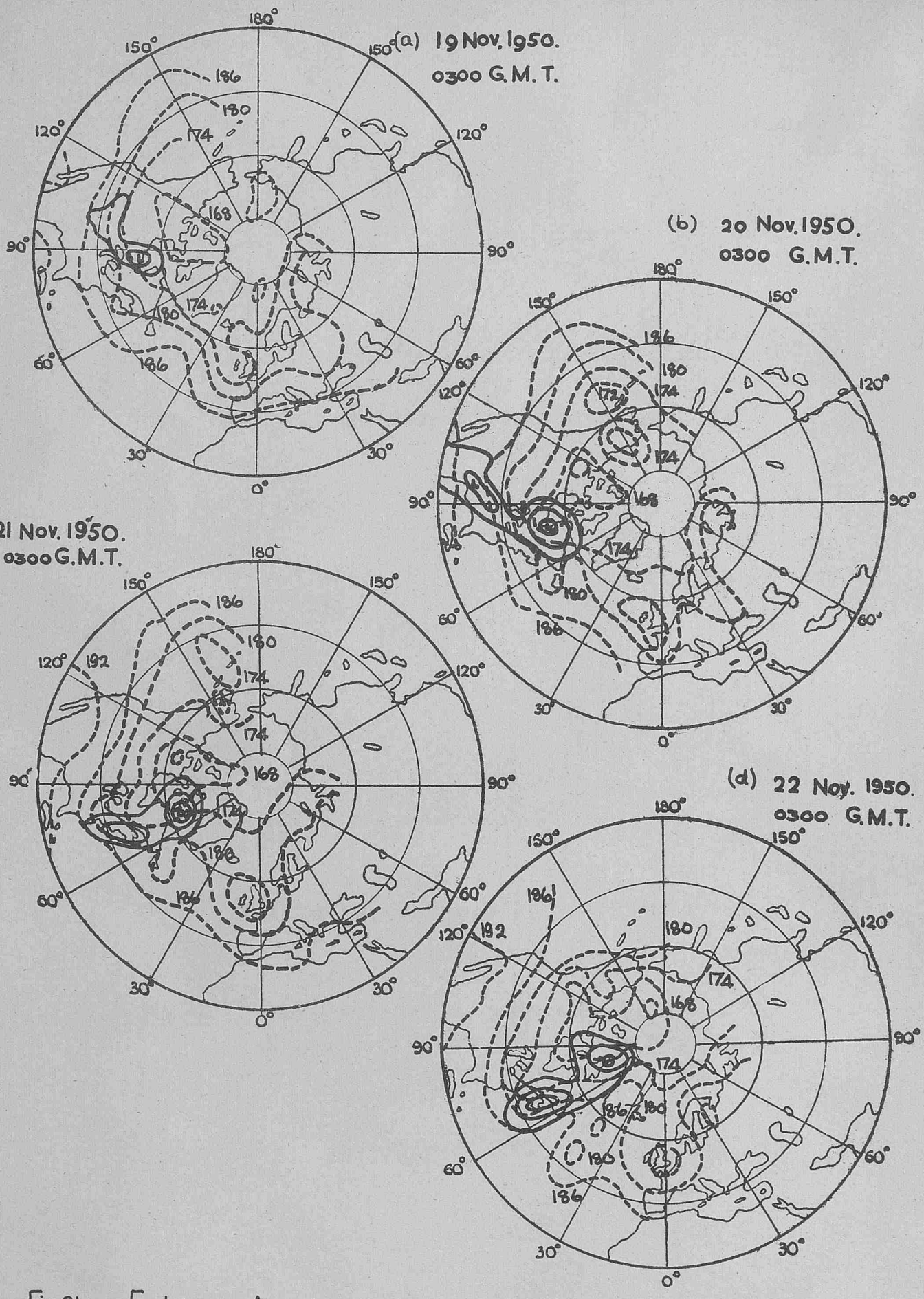
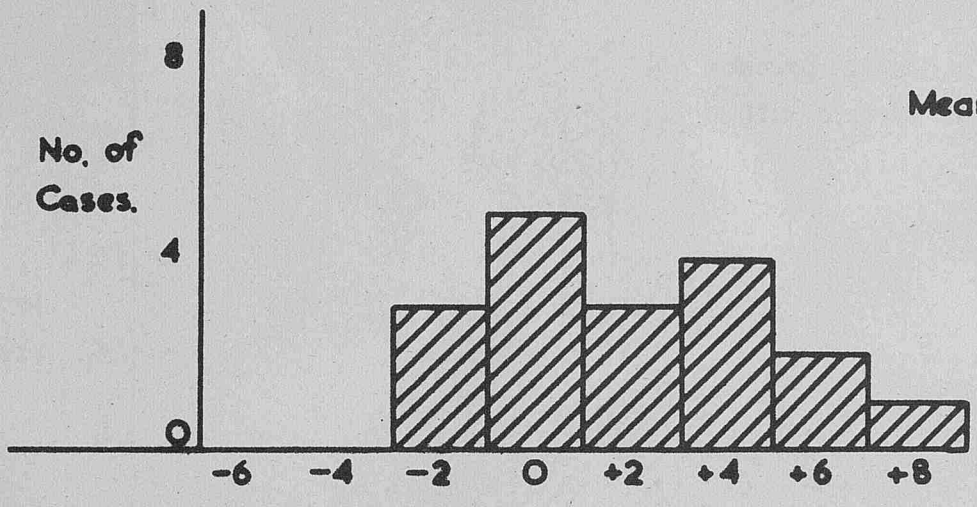
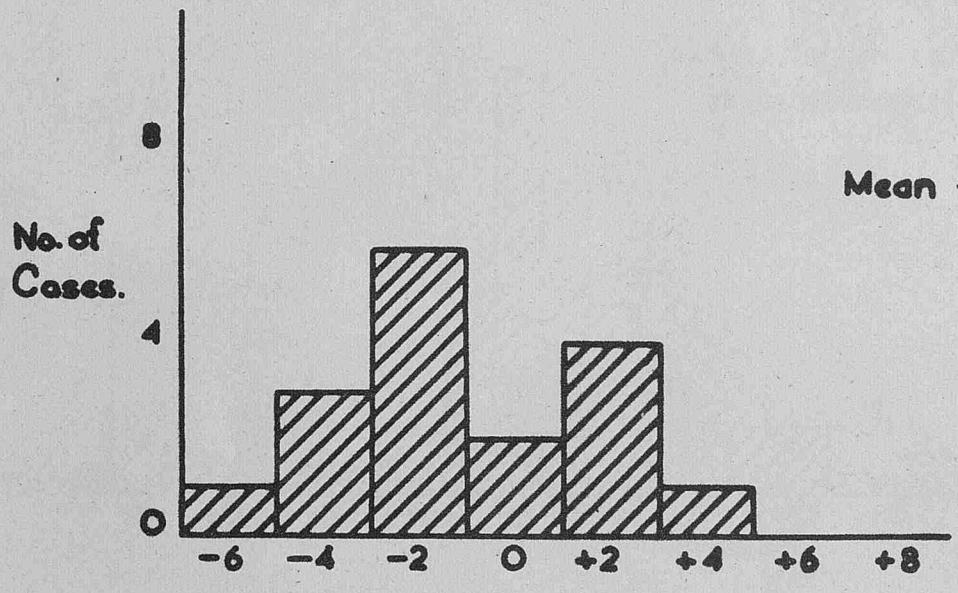


Fig. 21. Evolution of characteristic 500 mb. contour pattern during onset of NI spell 20 Nov. – 5 Dec 1950.
500mb. contours 0300 G.M.T. – – –, contour heights in hundreds of feet
Surface isobars (near area L) 0000 G.M.T. ———

Steering Deviation N class depressions



(a) Steering deviation in interval H_{-24} to H_{+24}



(b) Steering deviation in interval Ho to H_{+24}

Fig.22. Deviation between movement of low and direction of thickness lines over the centre.

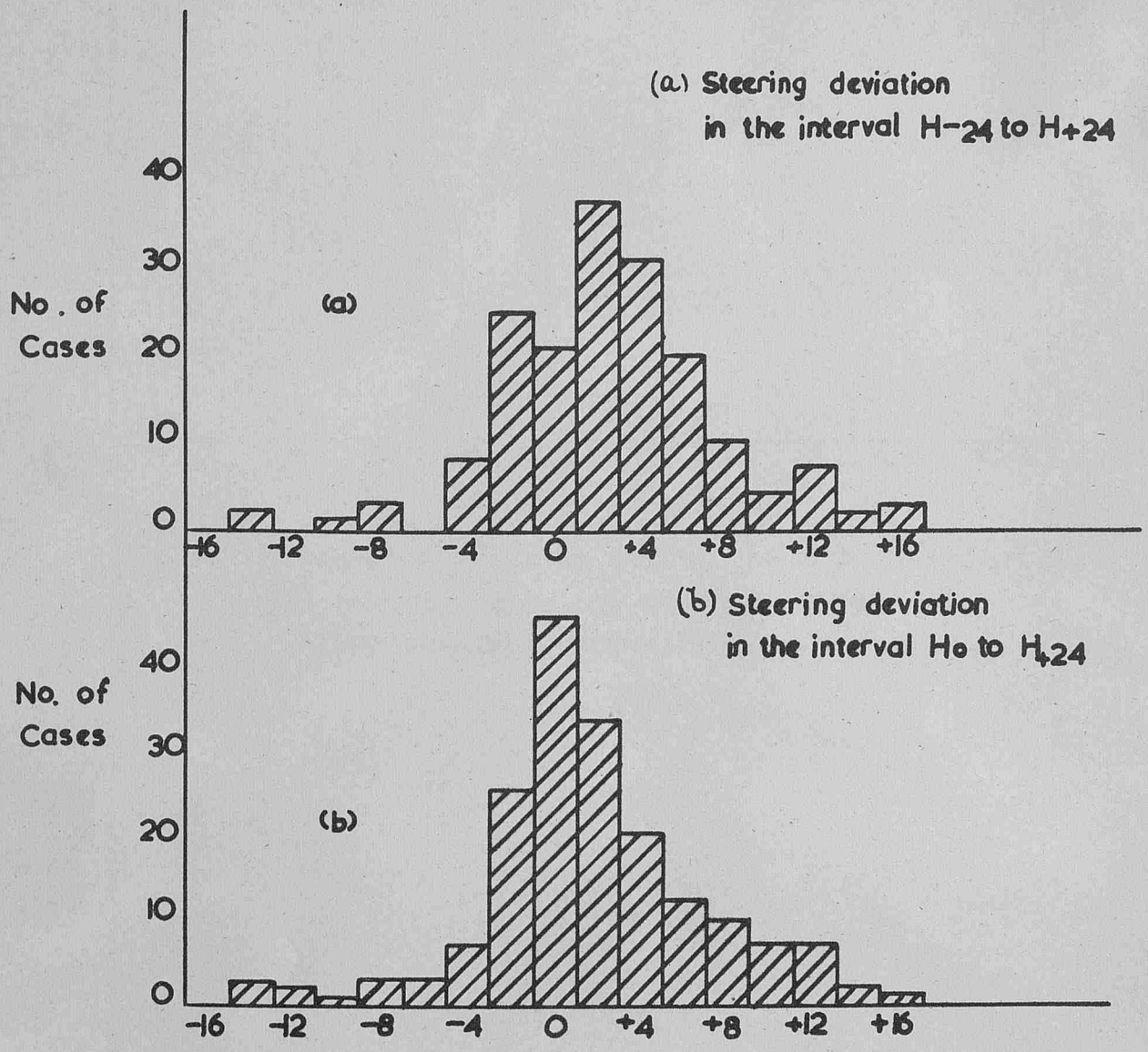


Fig.23. Deviation between movement of low and direction of thickness lines over the centre. Types C, E.

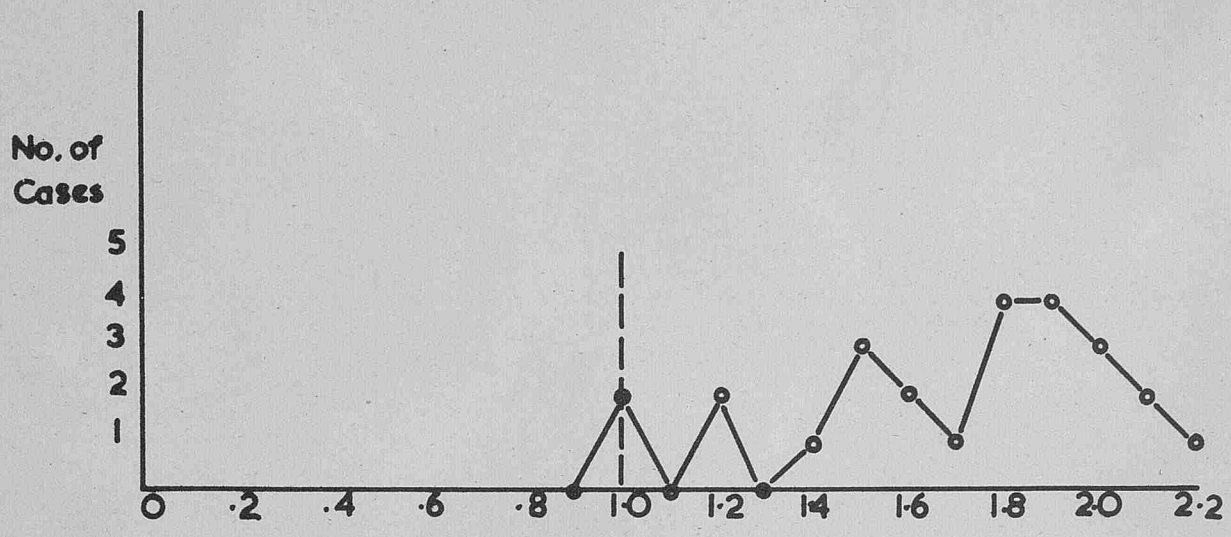


Fig. 24. Frequency of ratio actual thickness gradient to normal thickness gradient over the centres of rapidly deepening depressions.

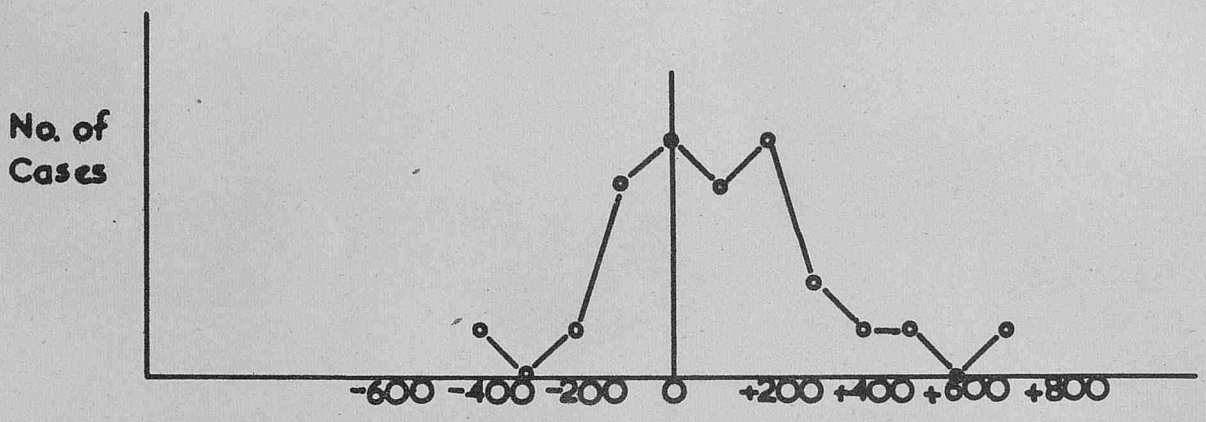


Fig. 25. Frequency of thickness anomaly over the centres of rapidly deepening depressions.

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A copy is available in the Library of the Meteorological Office."

M.R.P.778

S.C.II/135

4th December, 1952.

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METEOROLOGICAL RESEARCH COMMITTEE

WIND AT HIGH LEVELS OVER HONG KONG

By

R. F. M. HAY (Meteorological Office)

1. The observations

Radar-wind and radio sonde observations were begun at Hong Kong (Lat. 22°18'N. Long. 114°10'E.) in November, 1949. For most of the period observations were made on weekdays near 0800 Hong Kong local time. As Summer Time (G.M.T. + 9 hours) is used in Hong Kong from the first weekend of April to the last weekend of October this means that the observations refer to 0001Z during the summer period and to 2300Z for the remainder of the year. This note aims at a summary of the main features of the high level wind distribution and attention has been confined to the results for the two years 1950 and 1951.

2. General features

The principal features of the upper wind distribution at Hong Kong in the lower layers are well known, and are only briefly given here; the details are fully described elsewhere (1). The N.E. monsoon begins in September but does not become fully developed until a month later. At all times it is a shallow current, the mean depths in various months being October 8,000 ft., November 6,000 ft., December 5,000 ft., January and February 4,000 ft., March and April 3,000 ft.

Above this current there is a sharp change to westerlies usually marked by a strong inversion and a marked wind shear. These are the southern branch of the zonal N. Hemisphere westerlies which split on the western side of the Asian plateaux (2).

In April the N.E. monsoon weakens though occasionally 'bursts' occur even in May. With the northward movement of the inter-tropical convergence zone and the formation of monsoonal lows over N. India and China, the southerly monsoon begins in late May and is fully established from June to the end of August. As a rule winds in the southerly monsoon have an easterly component at least as often as a westerly component at low levels. The monsoon current is generally about 6,000 ft. deep with a gradual change to upper easterlies at higher levels but its depth fluctuates widely, it may be over 12,000 ft. deep, or on occasions even in summer, it may be absent. There is seldom a marked limit between the southerly current and the upper easterlies and on many days in summer a deep transition layer of light variable winds is found.

/3.

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3. Mean winds at high levels

The number of observations available at the higher levels is shown in the table below.

mb.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	1st Jan., 1950-31st Dec., 1951
60	1	2	0	3	7	8	9	11	11	19	6	2	79
100	2	3	2	10	22	24	31	28	26	33	25	7	213
150	6	9	2	18	31	32	42	35	32	36	33	20	296
200	19	17	14	26	34	38	46	38	36	37	43	30	378
300	45	35	26	36	37	41	50	39	41	44	50	48	492
850	49	41	39	36	38	36	47	40	42	48	51	49	516

Using all these observations vector and scalar mean winds were computed, the results are shown in Tables 1 and 2, and Figure 1. Strong westerlies at the 300 and 200 mb. levels show prominently in the months November to April and easterlies prevail above 300 mb. in July to September. Above 300 mb. the period of the summer in which easterlies prevail increases with height, until at 100 mb. they are found from June to September.

Frequency tables compiled for 200, 150 and 100 mb. levels are shown in Table 3 and Figure 4. Considering wind speeds alone the 20-30 kt. range is evidently most frequent at higher levels in summer. At these three levels winds in the 30-80 kt. ranges occur with almost equal frequency. In winter as would be expected the 200 mb. level shows a greater frequency of winds in range values above 50 kt. than either of the higher levels.

The frequency distribution of the easterlies and westerlies shows up rather more clearly in the second parts of Table 3 and Figure 4. At 200 mb. in summer an easterly component of 10 kt. is most frequent (20% in the 5-14 kt. range) but the incidence of the upper easterlies shows up in a second maximum (14%) in the 25-34 kt. range. At 150 mb. in summer their frequency is still higher, up to 14% in the 35-44 kt. range; in other respects the distribution of components is broadly similar to that at 200 mb. (Table 3). The winter distribution of the westerly component at 200 mb. shows a "normal" distribution with a maximum frequency (32%) in the 45-54 kt. range; the distribution at 300 mb. resembles it except that strong westerlies are more frequent at the lower level. The other curves give the frequency distribution of the meridional components at 200 mb. for the two seasons and show that in both seasons the distribution is roughly "normal", light northerly components being common in summer and light southerly components in winter. The inclusion of March and September in the winter and summer seasons respectively did not affect the frequency distribution of any of the components shown in Figure 4.

4. Transition from westerlies to easterlies

The daily observations have been plotted for Hong Kong for the periods May to October, 1950 and 1951 and the levels 200, 150 and 100 mb. The results are shown in Figures 2 and 3 in which wind directions and speeds are plotted as ordinates on separate graphs against a common time scale as abscissa. The transition from westerlies to easterlies at the 100 mb. level occurs in late May and is fairly abrupt, after which almost uninterrupted easterlies prevail until early or mid October. At 150 mb. and 200 mb. the transition to easterlies occurs a week or two later, also the easterlies end a few days earlier than at 100 mb. At both levels and particularly at 200 mb. breaks of a few days in the easterlies occurred on several occasions. At 300 mb. westerlies were almost as prevalent as easterlies in these months.

/The

The approximate dates of earliest beginning and latest ending of occurrence of a short spell (as distinct from isolated observations) of easterlies over the two years were:-

	Beginning	Ending
100 mb.	(23rd May)	24th October
150	6th June	16th October
200	4th June	20th September
300	4th June	21st September

Isolated cases of easterlies at these levels occurred outside these dates but in no case earlier than 1st May or later than 31st October. At the 60 mb. level the few observations available show that easterlies were confined to the months April to October. 64 out of 68 observations in these months showed easterlies; the other 4 observations of westerlies were all at the end of October.

5. Extreme winds

As the table of vector mean winds suggests the strongest winds were found at the 300 mb. and 200 mb. levels in the winter months. In fact these winds comprise the southern branch of the upper westerlies blowing around the Central Asian plateaux, the days of strongest wind at Hong Kong being associated with jet stream development further to east over active portions of the W. Pacific polar front, which, in the winter months, usually extends from just south of Hong Kong to the vicinity of Japan. Table 4 shows that the highest wind speeds occurred in most months near these levels and with westerlies in all months except June to September. In these months the monthly extremes occurred with easterlies. The table shows that the highest wind speed (268/123 kt.) occurred on 10th December, 1951 and the highest observed speed in the easterlies (040/078 kt.) was on 9th July, 1951. The maximum wind on 17th June, 1951 (199/086 kt.) is interesting. The Hong Kong monthly weather summary states that the period 14th-16th was one with lower pressure and stronger winds than usual for the time of year and on 17th a tropical depression formed over the sea to south of Hong Kong. The unusual direction of the strong wind in this ascent, doubtless associated with the formation of the tropical depression, is confirmed by similar readings found on two ascents at Kagoshima (Lat. 31°34N. Long. 130°33'E.) on the same date.

Table 2 shows that the strongest monthly vector mean winds in the easterlies are 43 kt. and 31kt. in July at the 60 mb. and 100 mb. levels. These figures are well below mean values for Aden at comparable levels; however, easterlies exceeding 50 kt. are found on a number of days in the summer months at Hong Kong at these levels.

Bibliography

(1) M.O.404b. Weather in the China Seas, Vol. II, pp. 186 - 193.

(2) B. W. Thompson "An essay on the general circulation of the atmosphere over S.E. Asia and the W. Pacific." Quart. J.R. met. Soc. October, 1951, p. 573.

TABLE 1 - VECTOR RESULTANT WINDS AT HONG KONG

LEVEL	JAN.	FEB.	MAR.	APR.	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	DEC.- FEB.	MAR.- MAY	JUNE- AUG.	SEPT.- NOV.	LEVEL
60 mb.	266/28 (1)	277/24 (2)	-	152/07 (3)	095/16 (7)	067/29 (8)	074/43 (9)	079/35 (11)	085/19 (11)	091/04 (19)	272/11 (6)	276/29 (2)	274/27 (5)	102/13 (10)	075/35 (28)	086/06 (36)	60 mb.
100 mb.	266/44 (2)	261/25 (3)	249/15 (2)	263/21 (10)	278/15 (22)	047/18 (24)	058/31 (31)	070/26 (28)	053/11 (26)	288/05 (33)	250/22 (25)	255/51 (7)	258/43 (12)	271/16 (34)	060/25 (83)	274/05 (84)	100 mb.
150 mb.	252/50 (6)	256/41 (9)	273/20 (2)	261/33 (18)	274/31 (31)	359/15 (32)	055/28 (42)	068/13 (35)	063/06 (32)	289/15 (36)	254/35 (33)	248/53 (20)	250/50 (35)	269/31 (51)	046/18 (109)	268/14 (101)	150 mb.
200 mb.	253/63 (19)	255/48 (17)	254/56 (14)	272/41 (26)	272/33 (34)	336/06 (38)	057/19 (46)	069/12 (38)	030/06 (36)	297/16 (37)	262/38 (43)	256/55 (30)	255/55 (66)	267/40 (74)	052/11 (122)	276/18 (116)	200 mb.
300 mb.	254/61 (45)	264/55 (35)	254/60 (26)	267/43 (36)	268/27 (37)	292/04 (41)	075/09 (50)	077/06 (39)	032/03 (41)	273/12 (44)	264/35 (50)	256/53 (48)	257/56 (128)	263/41 (99)	065/04 (130)	269/16 (135)	300 mb.
500 mb.	257/38 (49)	268/43 (41)	262/39 (37)	257/31 (39)	265/17 (39)	237/07 (39)	134/04 (50)	217/02 (43)	106/04 (43)	279/05 (46)	259/22 (52)	260/35 (48)	262/38 (138)	261/29 (115)	205/03 (132)	258/08 (141)	500 mb.
700 mb.	257/19 (49)	260/19 (41)	267/22 (39)	251/22 (37)	249/11 (38)	223/10 (38)	193/08 (48)	217/07 (42)	109/05 (41)	041/03 (48)	245/09 (52)	272/14 (48)	262/17 (138)	257/18 (114)	210/08 (128)	225/02 (141)	700 mb.
850 mb.	179/02 (49)	201/09 (41)	218/09 (39)	210/13 (36)	184/07 (38)	197/08 (36)	181/07 (47)	210/06 (40)	094/08 (42)	072/13 (48)	087/07 (51)	073/05 (49)	166/03 (139)	205/09 (113)	194/07 (123)	082/09 (141)	850 mb.

Explanation:- 257/38 ... wind direction 257° true, speed 38 knots.
(49) ... figure below in brackets denotes the number of observations.
- ... dash denotes no observations.

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TABLE 2 - AVERAGE SCALAR WIND SPEED (KNOTS) AT HONG KONG

LEVEL	JAN.	FEB.	MAR.	APR.	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	DEC.- FEB.	MAR.- MAY	JUNE- AUG.	SEPT.- NOV.
mb. 60	28 (1)	23 (2)	-	7 (3)	16 (7)	30 (8)	43 (9)	36 (11)	22 (11)	13 (19)	14 (6)	29 (2)	27 (5)	14 (10)	37 (28)	16 (36)
100	45 (2)	25 (3)	16 (2)	22 (10)	21 (22)	23 (24)	33 (31)	27 (28)	21 (26)	17 (33)	27 (25)	51 (7)	44 (12)	21 (34)	28 (83)	21 (84)
150	34 (6)	40 (9)	20 (2)	35 (18)	35 (31)	25 (32)	33 (42)	21 (35)	18 (32)	24 (36)	38 (33)	55 (20)	48 (35)	35 (51)	27 (109)	27 (101)
200	65 (19)	49 (17)	57 (14)	43 (26)	36 (34)	24 (38)	26 (46)	19 (38)	20 (36)	27 (37)	41 (43)	57 (30)	57 (66)	43 (74)	23 (122)	30 (116)
300	63 (45)	57 (35)	62 (26)	46 (36)	29 (37)	15 (41)	15 (50)	14 (39)	14 (41)	24 (44)	37 (50)	55 (48)	58 (128)	44 (99)	15 (130)	26 (135)
500	40 (49)	46 (41)	41 (37)	33 (39)	18 (39)	13 (39)	11 (50)	12 (43)	11 (43)	15 (46)	25 (52)	36 (48)	40 (138)	30 (115)	12 (132)	18 (141)
700	22 (49)	23 (41)	23 (39)	24 (37)	15 (38)	15 (38)	13 (48)	13 (42)	12 (41)	13 (48)	13 (52)	18 (48)	21 (138)	21 (114)	14 (128)	13 (141)
850	11 (49)	14 (41)	13 (39)	15 (36)	14 (38)	15 (36)	14 (47)	13 (40)	13 (42)	16 (48)	12 (51)	12 (49)	12 (139)	14 (113)	14 (123)	14 (141)

Figures in brackets give number of observations.

		20	Number of observations
WINTER (Dec. Jan. Feb.)	100 m		12
	150 m		35
	200 m		66
SUMMER (June, July, Aug.)	100 m		83
	150 m		109
	200 m		122

		knots -125 -116	+25 +34	+35 +44	+45 +54	+55 +64	Number of observations
WINTER (Dec.-Feb.) 200 mb.	N-S	-	2	-	-	-	66
	E-W	1	-	-	-	-	66
WINTER (Dec.-Feb.) 300 mb.	N-S	-	1	-	-	-	128
	E-W	-	-	-	-	-	128
SUMMER (June-Aug.) 150 mb.	N-S	-	15	3	1	1	109
	E-W	-	12	14	5	1	109
SUMMER (June-Aug.) 200 mb.	N-S	-	6	2	-	-	122
	E-W	-	14	10	3	-	122

TABLE 4 - EXTREME VALUES OF UPPER WINDS AT HONG KONG

Year	Month	Day	Wind ^{o}true	Wind kt.	Level of occurrence mb.	Level of termination of ascent mb.	Number of ascents in given month.
1950	Jan.	14	265	119	200	200	24
	Feb.	4	245	99	425	300	18
	Mar.	17	235	96	241	200	19
	Apr.	5	290	73	216	(200)	18
	May	7	270	58	252	100	15
	June	30	080	54	89	60	20
	July	14	065	72	153	100	26
	Aug.	7	085	54	65	60	24
	Sept.	1	215	60	186	100	21
	Oct.	31	310	55	218	60	22
	Nov.	30	285	100	233	(200)	25
	Dec.	4	285	87	342	300	22
1951	Jan.	13	286	111	339	(300)	25
	Feb.	19	286	92	323	300	23
	Mar.	17	250	102	310	300	20
	Apr.	3	289	99	250	(200)	24
	May	(25	286	69	179	150)	26
		(31	265	69	145	150)	
	June	17	199	86	213	200	22
	July	9	040	78	131	(100)	27
	Aug.	26	064	66	189	(150)	24
	Sept.	11	061	55	75	40	25
	Oct.	29	220	99	263	(200)	29
	Nov.	20	250	81	308	200	27
	Dec.	10	268	123	64	(60)	28

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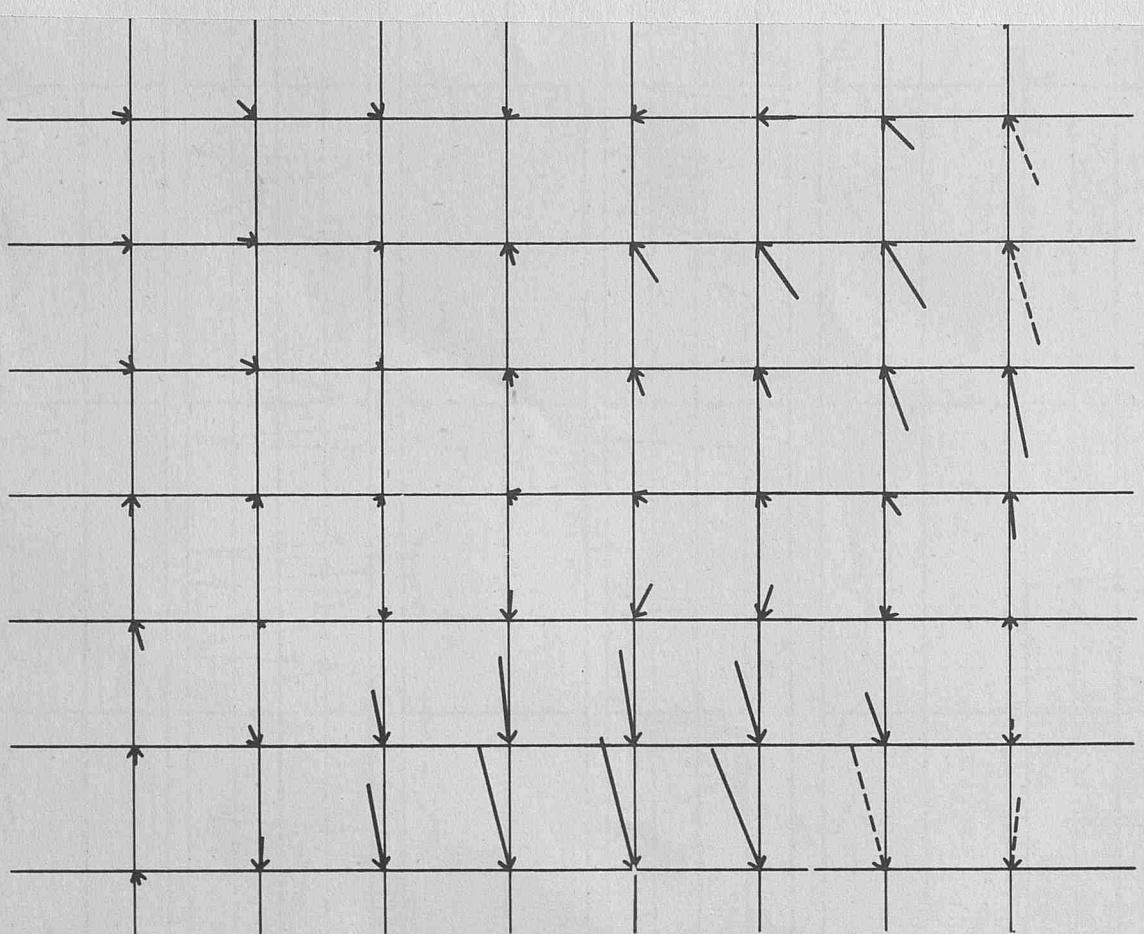
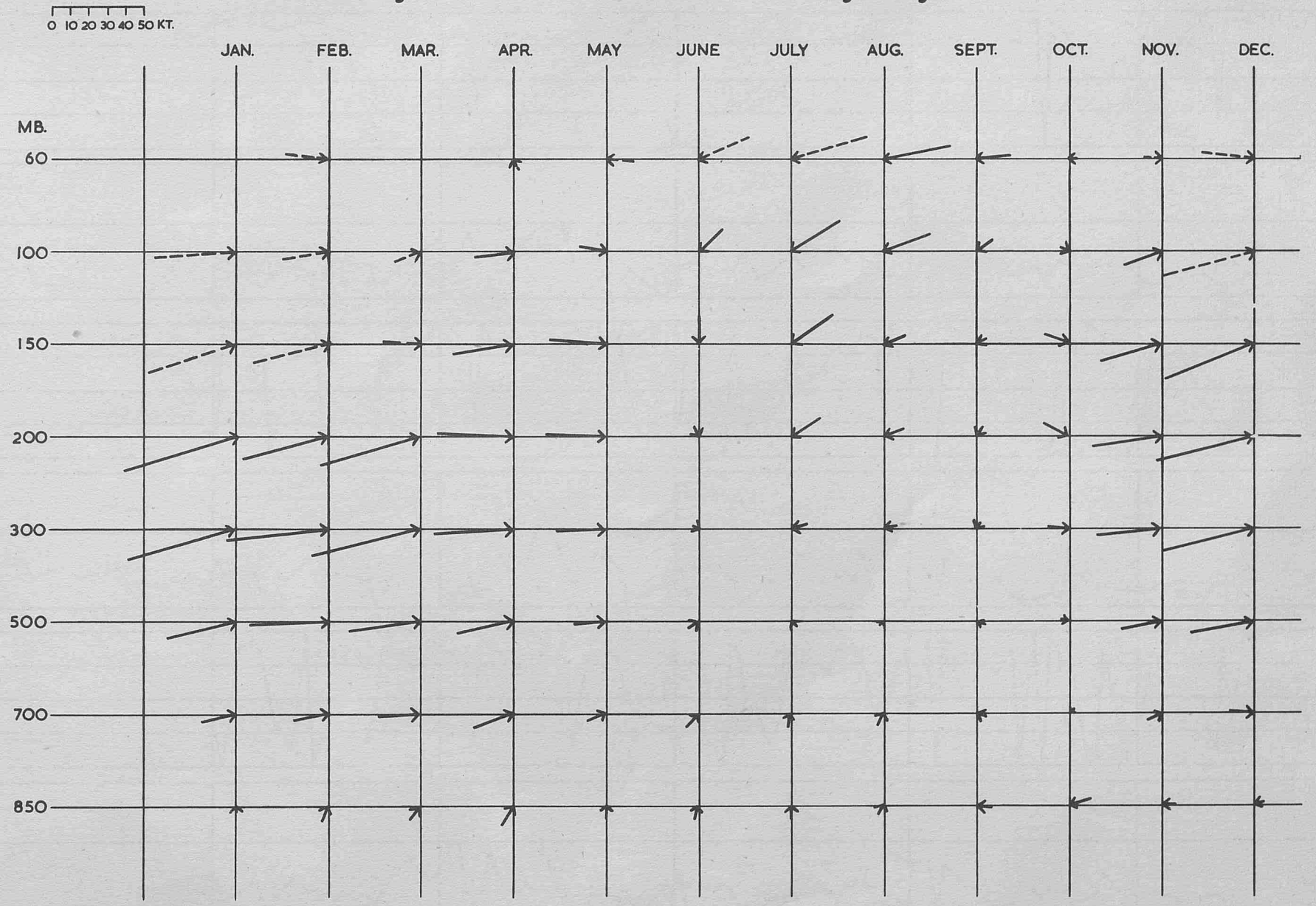


t winds at Hong Kong.

M.R.P. 778

d lines denote means of <10 observations)

Fig.1. Vector resultant winds at Hong Kong.



Computed from all observations. (dotted lines denote means of <10 observations)

Figure 2.—INCIDENCE OF UPPER EASTERLIES AT HONG KONG, 1950

Figure 3.—INCIDENCE OF UPPER EASTERLIES AT HONG KONG, 1951

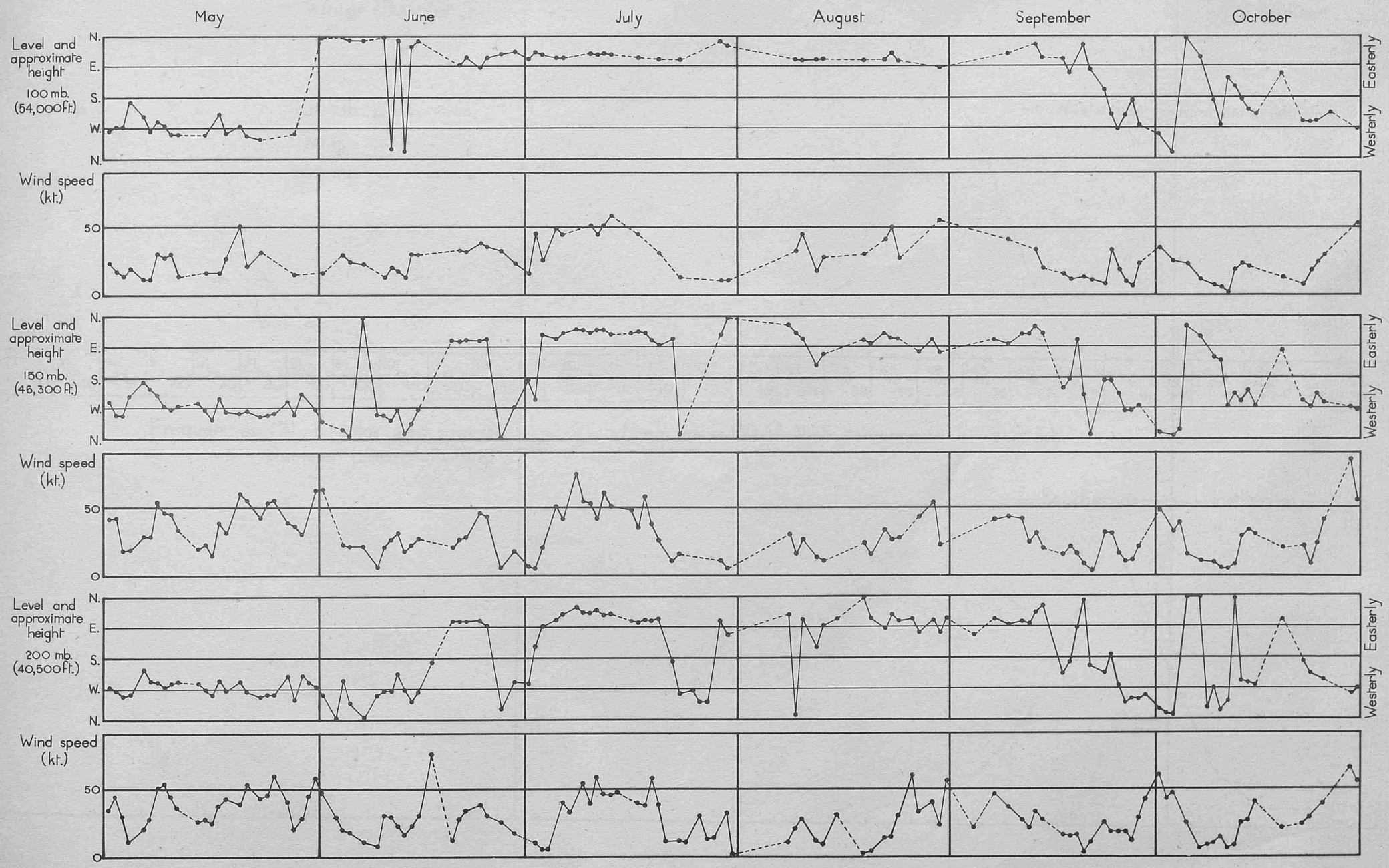
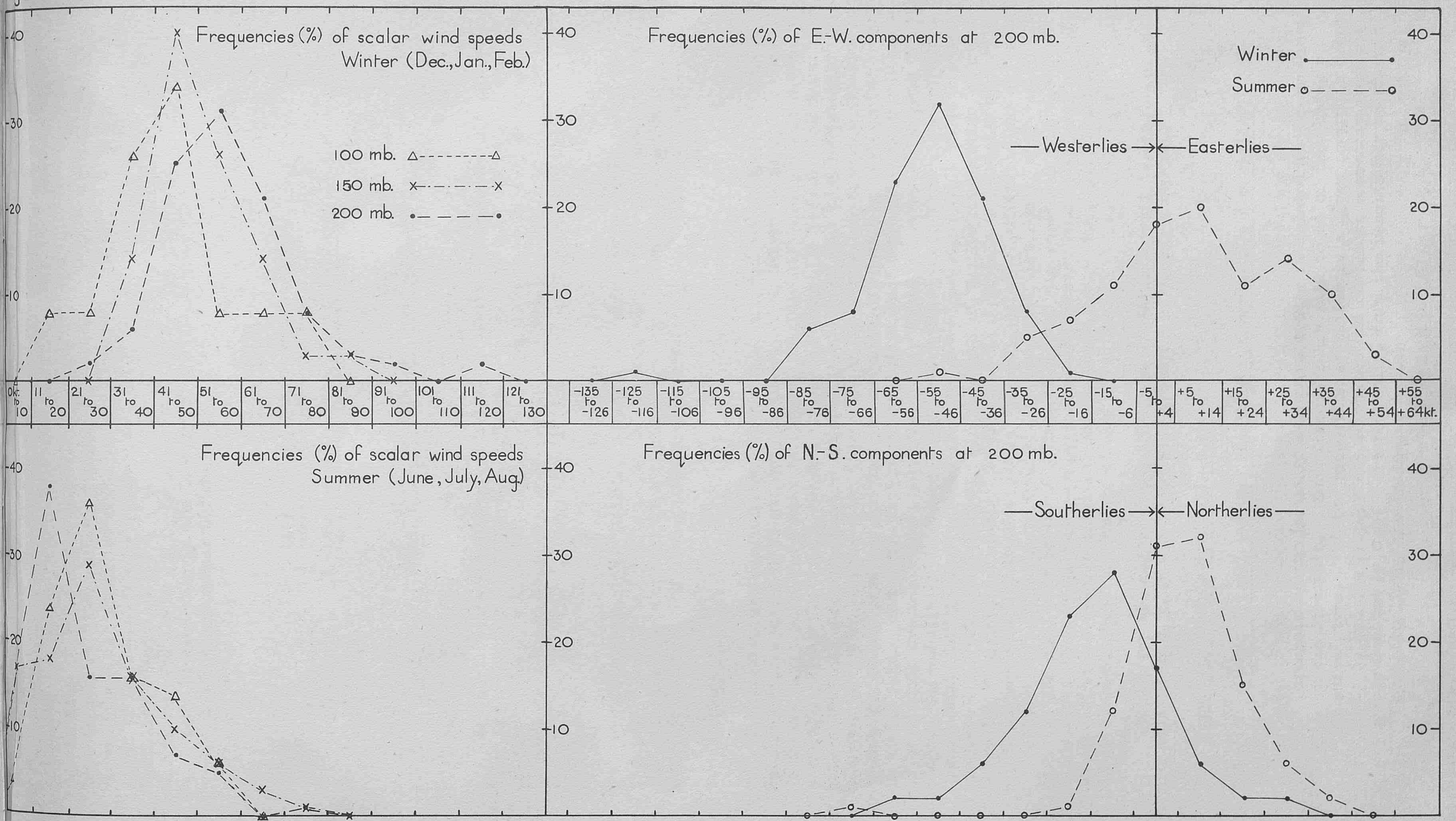


Figure 4.


Frequencies (%) of scalar wind speeds
Winter (Dec., Jan., Feb.)
100 mb.
150 mb.
200 mb.
0kt. to 10
11 to 20
21 to 30
31 to 40
41 to 50
51 to 60
61 to 70
71 to 80
81 to 90
91 to 100
101 to 110
111 to 120
121 to 130
Frequencies (%) of E.-W. components at 200 mb.
Winter
Summer
Westerlies
Easterlies
-135 to -126
-125 to -116
-115 to -106
-105 to -96
-95 to -86
-85 to -76
-75 to -66
-65 to -56
-55 to -46
-45 to -36
-35 to -26
-25 to -16
-15 to -6
-5 to +4
+5 to +14
+15 to +24
+25 to +34
+35 to +44
+45 to +54
+55 to +64kt.
Frequencies (%) of scalar wind speeds
Summer (June, July, Aug.)
Frequencies (%) of N.-S. components at 200 mb.
Southerlies
Northerlies
40
30
20
10

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 788 by Mr. A.F. Crossley. A copy is available in the Library of the Meteorological Office."

S.C.II/138
24th March, 1953.

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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Measures of success in forecasting

by

A.F. Crossley (Meteorological Office)

Summary.

An index of the usefulness of forecasts of the 'black or white' type is adopted in the form $I = \frac{1}{2}(c+c')$, where c, c' are the forecast accuracies of blacks and whites respectively. The success of this type of forecast, defined as the improvement beyond the point reached by persistence, is given by $\mathcal{I} = (I-I_o)/(1-I_o)$, where I_o is the value of I corresponding with persistence forecasts.

For forecasts of continuously varying elements, Priestley's index is adopted, $P = (1-\varepsilon^2/\sigma^2)^{\frac{1}{2}}$, where ε^2 is the variance of the forecast errors and σ^2 the time variance of the element. This index measures the improvement over persistence, the latter corresponding with P = 0.

1. Introduction.

Information on the accuracy of forecasts is required in order to provide a measure of progress and to compare the merits of different techniques. While it is now amply recognised that in general 100 per cent accuracy can never be achieved (Sutcliffe, 1952), there is nevertheless still room for progress in many directions and considerable effort is being expended to make advances wherever possible. It is then important to be provided with a quantitative assessment of the progress, or lack of progress, achieved by any developments in techniques. Since in practice it is seldom that more than one technique is applied to precisely the same set of data, it is necessary to be able to compare the success of different techniques applied to different sets of data. Hence suitable indices of success need to be devised which will be as far as possible independent of the technique used and of the varying circumstances in which the forecasts are made. Further, it is desirable to minimise the effects of persistence in the elements being forecast, since repetitive cases can often be forecast without much aid from highly developed techniques, with the result that the presence of persistence tends to conceal any essential usefulness of those techniques. In other words, a correct forecast of a substantial change in conditions is usually reckoned as being more meritorious than a correct forecast when conditions remain steady.

Indices for forecasts of the 'black or white' type will be considered first, and subsequently those for forecasts of scalar or vector elements.

2. Forecasts of 'black or white' type.

2.1. Useful effort, and the condition for worthwhile forecasting.

In an earlier paper (Crossley, 1952) it was shown how the usefulness of forecasts of two mutually exclusive events is related to the accuracy of the forecasts. Such forecasts are said to be of the 'black or white' type. If the proportion of whites in a given period is denoted by b, and the proportions of whites and of blacks correctly forecast are denoted by c, c' respectively, then the results may be shown symbolically by the following table:-

D 70024
ENCLOSURE 3 TO ER-1664-54 AF630120

/Table I

Table I - Forecast contingency table

	Forecast	Not forecast	Total
White	bc	$b(1-c)$	b
Black	$(1-b)c'$	$(1-b)(1-c')$	$1-b$
Total	$bc+(1-b)c'$	$b(1-c)+(1-b)(1-c')$	1

The useful effort, E, defined as the successful proportion of forecasts of white, is given by

$$E = \frac{bc}{bc+(1-b)(1-c')} \qquad (1)$$

while a similar expression defines the useful effort in regard to forecasts of black. In this connection, one desires to know whether, and to what extent, forecasting constitutes an improvement on a policy of merely waiting for suitable opportunities to come along. If forecasting is to be worth while in this respect, it is necessary that the number of successful forecasts in a given period should exceed the number of occurrences of the event; that is, one requires E to exceed b. The value of E in relation to b may therefore be taken as a measure of the 'usefulness', but this leads to certain difficulties to which Mr. P.M. Shaw and Mr. F.E. Lumb have drawn attention. Thus the value of E becomes unity when $c' = 1$, whatever the value of c; and it becomes zero when $c = 0$, whatever the value of c'. Moreover, the possible range of both E and E/b varies with b, so that two or more values of either index are not readily comparable. An alternative condition for worthwhile forecasting was given in the earlier paper, where it was seen that the condition $E > b$ leads, after reduction, to

$$c+c' > 1 \qquad (2)$$

which shows that the sum of the two forecast accuracies must exceed unity for worthwhile forecasting of whites. This condition is however symmetrical with regard to both c and c', so that the same inequality follows from consideration of the useful effort in regard to the black occasions. The value of $c+c'$, or some function of this expression, may therefore be considered as a more appropriate index of the usefulness of the forecasts. Various other indices may also be derived to express the same condition, and their respective merits will now be considered.

2.2. Types of index of usefulness.

When the results of a set of forecasts are summarised, the contingency table takes the following form:-

Table II - Summary of results of forecasts

	Forecast	Not forecast	Table
White	A	B	A + B
Black	C	D	C + D
Total	A + C	B + D	A+B+C+D

From this we have, if E relates to the forecasts of whites,

$$E = \frac{A}{A+D}, \qquad b = \frac{A+B}{A+B+C+D},$$

$$c = \frac{A}{A+B}, \qquad c' = \frac{C}{C+D}$$

/The

The condition for worthwhile forecasting, $E > b$, then becomes

$$\frac{A}{A+D} > \frac{A+B}{A+B+C+D}$$

This leads to an index in the form

$$I_2 = \frac{A(A+B+C+D)}{(A+D)(A+B)} \qquad \text{.....(3)}$$

which also will exceed unity for worthwhile forecasting. Although both are derived from the same inequality, the numerical value of I_2 is in general different from that of $c+c'$. Either of the inequalities (2) and (3), when reduced to their simplest form in terms of A,B,C,D, leads to

$$AC > BD$$

and this gives yet another possible index,

$$I_3 = \frac{AC}{BD} \qquad \text{.....(4)}$$

One may, in fact, derive an unlimited number of indices of this type, since from any inequality a new one may be formed by adding the same arbitrary quantity to both sides. The problem then is to choose the one most suitable for the purpose under consideration. The general characteristics of such indices will be appreciated from a discussion of the three given above.

Let us consider the range of variation of each of these indices. It may be shown that each of them increases whenever c or c' increase. The lower limits of the indices therefore occur when the forecast is wrong every time:- $c = c' = 0$, or $A = C = 0$; the upper limits arise when the forecast is correct every time:- $c = c' = 1$, or $B = D = 0$. The following are the ranges of variation:-

$$I_1, \quad 0 \leqslant c+c' \leqslant 2$$

$$I_2, \quad 0 \leqslant \frac{A(A+B+C+D)}{(A+D)(A+B)} \leqslant 1 + \frac{C}{A}$$

$$I_3, \quad 0 \leqslant \frac{AC}{BD} \leqslant \infty$$

Of these ranges, the first is constant, the second varies with the data, and the third is infinite. In terms of b, c, c' we have

$$I_2 = \frac{c}{bc + (1-b)(1-c')}, \qquad \text{.....(5)}$$

so that the value of the index depends on the frequency of occurrence of the element which is being forecast. Thus for two different periods, different values of the index may be obtained even though the accuracy of forecasting is the same in both cases. Moreover the symmetry between forecasts of blacks and whites is lost with this index, and the expressions (3) or (5) derived from the forecasts of white are not identical with the corresponding expressions for the blacks. For these reasons, I_2 does not constitute a satisfactory index.

Such considerations suggest that the most suitable index should be a symmetrical function of c and c', and independent of b. This is true of I_1 and also of I_3, which can be expressed as:-

$$I_3 = \frac{cc'}{(1-c)(1-c')},$$

Of the two, clearly I_1 is to be preferred on account of its simplicity, its linear form, and its finite range. Since however it is more convenient that the range should run from 0 to 1, the index will be adopted in the form:-

$$I = \tfrac{1}{2}(c+c') \qquad \text{.....(6)}$$

It is then the mean of the two forecast accuracies; for worthwhile forecasting of either whites or blacks, in the sense used here, its value must exceed one-half, or 50 per cent. It will be termed the "usefulness". In terms of the elements of Table II, it may be expressed in the form:-

$$I = \frac{1}{2}\left(\frac{A}{A+B} + \frac{C}{C+D}\right) \qquad \text{.....(7)}$$

While this index therefore gives a suitable measure of the usefulness of the forecasts, the value is in part dependent on persistence. It is necessary to proceed a stage further in order to obtain a measure of the success of forecasting, which may be defined as the degree of improvement beyond the point reached by persistence. This will be considered in the next section.

2.3. Elimination of the effects of persistence.

We next consider how estimates of success in forecasting are related to persistence or repetition of the element concerned. In the extreme case, if the element were known to be a constant, no skill would be required once that fact had been observed. In ordinary cases, there is a repetitive tendency in the element which is reflected in a tendency to forecast a recurrence of the event. Often this does not require much skill and is more or less independent of the particular technique used; on the other hand, where a change is expected but the time of occurrence is uncertain, much care may be exercised before a forecast of no change is issued. Generally, however, persistence is a factor which operates towards accurate forecasting irrespective of the techniques used. In assessing the value of a technique, it is therefore desirable that the effects of persistence should be removed.

Suppose then that the value of the index I or $\tfrac{1}{2}(c+c')$ has been obtained for a particular set of black and white forecasts. Let us now similarly compute another index I_o for the same occasions on the assumption that the forecast every time is one of repetition of the existing conditions. Then the difference $I - I_o$ would give a measure of the success of the technique used, over and above that to be obtained by use of persistence. To be more precise, we need to consider the improvement in the value of I over that of I_o in relation to the maximum possible improvement. Since the maximum of either index is unity, this leads to the following expression* for the measure of success of a forecasting technique,

$$J = \frac{I - I_o}{1 - I_o} \qquad \text{.....(8)}$$

* This expression resembles one proposed by Meetham (1947), in which an arbitrary marking system was used for assessing terms corresponding with I and I_o.

If I exceeds I_o, the value of J lies between 0 and 1. If I is less than I_o, then J is negative and may take any value; in this case the forecasting technique is worthless, since the elementary method of persistence gives better results.

/The

The particular form in which forecasts of persistence are made will depend on the nature of the element being forecast. Forecasts of no change from current conditions will be appropriate in certain circumstances; if on the other hand there is a marked diurnal variation, a forecast of repetition after 24 hours may be more appropriate. Again, if the element is largely influenced by advection, the persistence forecast might consist of the value of the element in a location displaced up-wind by the time-interval of the forecast.

We see then that the index J, eqn. (8), measures the success in forecasting beyond that which would be achieved by the application of persistence. Moreover, in virtue of the definitions of I and I_o, the forecast accuracies of both white and black occasions are treated on an equal footing, while any effects due to the frequency of occurrence of the element considered are eliminated. As regards one set of data alone, the value of the difference $I - I_o$ would be a sufficient indication, but in comparing different sets or different techniques the expression J should be used to evaluate the success.

2.4. Applications to black and white forecasts.

2.4.1. Forecasts of change of weather type made in a series of 4-day forecast trials have been discussed previously (Crossley, 1952). The accuracy of forecasts of a change of type (c) was found to be 0.53, the accuracy of forecasts of no change (c') 0.82, hence I = 0.675. If one had forecasted no change of type, i.e. persistence, on every occasion, the result would have been c = 0, c' = 1, I_o = 0.5. This again illustrates that I must exceed 0.5 for the results to be worthwhile. In this case the success over persistence forecasting is given by J = 0.175/0.5 = 0.35.

2.4.2. Formation of secondary depressions.

Another example which was discussed previously (loc. cit.) concerns the application of Sawyer's criterion for the formation of secondary depressions at points of occlusion. The values were c = 0.82 for the formation, and c' = 0.86 for the non-formation of a secondary. These give I = 0.84. In this example, the question of persistence is irrelevant.

2.4.3. Forecasts of frost on conductor rails.

The results of certain forecasts of the temperatures of conductor rails were supplied by Mr. F.E. Lumb, Meteorological Office, Preston, from which the following table has been prepared.

Table III - Forecasts of conductor rail temperatures

Temperature	Forecast	Not forecast	Total
32°F or below	33	8	41
Above 32°F	112	13	125
Total	145	21	166

This gives c = 0.82 for occurrence of frost, c' = 0.90 for its non-occurrence, and I = 0.85. No information is available in this case regarding persistence.

2.4.4. Forecasts of rain.

An analysis of forecasts of rain contained in broadcasts by the B.B.C. at 5.55 p.m. each day during a period of 12 months in 1946-47 is given by Gold (1947). The parts of the forecasts concerning rain in south east England within the ensuing 24 hours were checked by comparison with observations at Kew Observatory.

The 267 unqualified forecasts of rain or no rain are summarised in Table IV.

Table IV - Forecasts of rain in S.E. England

	Forecast	Not forecast	Total
Rain	167	21	188
No rain	59	20	79
Total	226	41	267

For the accuracy of forecasts of rain, c = 0.89, and of no rain, c' = 0.75, giving I = 0.82, a high figure. Information is not given from which the corresponding value for persistence forecasting could be obtained, but the result would clearly be quite low.

2.4.5. Forecasting the fog-point.

Corby and Saunders (1952) give the results of a test of a method of predicting the temperature at which radiation fog forms. On potentially foggy occasions during the period August 1st to October 5th, 1950 the fog-point on the ensuing night was estimated from upper air soundings at 1500 G.M.T. by a method due to Saunders, and the results were compared with observations at a number of stations. The results were separated into two groups according to whether fog did or did not form. In the fog group, the forecast was considered successful if the predicted fog-point did not exceed the actual fog-point by more than $1^{o}F$; in the other group, the forecast was considered successful if the predicted fog-point was less than the night minimum temperature. Table V summarises the findings.

Table V - Forecasts of the fog-point

	Forecast	Not forecast	Total
Fog	44	11	55
No fog	78	15	93
Total	122	26	148

In this example we are concerned with forecasts of one type of event in two different sets of conditions, but at the time of the forecast it is not known which set applies. Since c = 0.80, c' = 0.84, the results show that the forecast accuracy is practically independent of the subsequent occurrence of fog. Interpreted as forecasts of fog, the usefulness is I = 0.82, but this is an overestimate, since in practice it is necessary also to forecast the night minimum temperature.

2.4.6. Forecasts of visibility at Northolt.

Although visibility is an element subject to continuous variation, the forecasts are conveniently considered under the head of 'black or white', since the position of an observation on the conventional visibility scale is usually of more importance than the actual visibility distance. Table VI contains results in regard to forecasts of the local visibility above and below certain limits; they were made four times daily between November, 1951 and March, 1952, the total number of forecasts being 556.

/Table VI

Table VI - Forecasts of visibility at Northolt
(November, 1951 - March, 1952)

Visibility limit	Forecast period								
	1 hr.			4½ hr.			12 hr.		
	c	c'	I	c	c'	I	c	c'	I
220 yd.	100	71	85	99	42	71	99	7	53
550 yd.	99	76	87	98	60	79	99	32	65
1,100 yd.	99	90	95	96	61	79	95	44	69
2,200 yd.	95	90	93	91	73	82	88	63	75
4,400 yd.	92	89	91	87	76	81	78	64	71
6¼ mi.	68	99	83	37	96	67	13	95	54

Note:
c, c' are the percentage accuracies of forecasts above, below the limits indicated, and $I = \frac{1}{2}(c+c')$ is the percentage usefulness.

Among the several points of interest to be inferred from this table, the following are mentioned:-

(i) The forecast accuracies c and c', as well as the index I, decrease as the period of the forecast increases. This is only to be expected.

(ii) The values of c decrease, and the values of c' increase, as the visibility limit increases. This too is to be expected.

(iii) For each forecast period, the value of I is greatest for moderate limits, and least for the extreme limits. This shows that the forecasting is most useful in regard to the moderate limits with a maximum usefulness at about 2,200 yd., and that it becomes less useful as the limits become more extreme. In particular, forecasts for 12 hours ahead for the limits of 220 yd. and 6¼ mi. are only just within the useful range, for which I must exceed 50 per cent. This is due almost entirely to the very low accuracy of forecasting visibility less than 220 yd. or greater than 6¼ mi., for a 12 hr. period.

The results which would have been obtained by persistence forecasting, i.e. by making the forecast visibility always the same as the actual visibility at the time of issue, are shown in Table VII.

Table VII - Persistence forecasts of visibility
at Northolt
(November, 1951 - March, 1952)

Visibility limit	Forecast period								
	1 hr.			4½ hr.			12 hr.		
	c_o	c_o'	I_o	c_o	c_o'	I_o	c_o	c_o'	I_o
220 yd.	100	71	85	98	33	65	98	29	63
550 yd.	99	86	93	97	48	73	97	24	61
1,100 yd.	98	89	93	97	63	80	95	40	67
2,200 yd.	96	85	91	91	55	73	89	47	68
4,400 yd.	90	89	89	80	73	77	70	55	63
6¼ mi.	85	97	91	52	91	71	30	84	57

(For explanation see note to Table VI)

Comparison of Tables VI and VII shows: (i) that for forecasts of visibility greater than a given limit, there is little difference between c and c_o except at 6¼ mi., where persistence gives definitely better results; (ii) that for forecasts below a given limit, c' is little different from c'_o for 1 hr. ahead, but rather better at 4½ hr. and 12 hr. with the exception of 220 yd. at 12 hr., where persistence is the better.

The usefulness as indicated by I or I_o can be similarly compared, but for possible comparison with other data, we compute the index of success (J), the values of which are shown in Table VIII.

Table VIII - Percentage success of visibility forecasting at Northolt (November, 1951 - March, 1952)

Visibility limit	Forecast period		
	1 hr.	4½ hr.	12 hr.
220 yd.	0	15	-29
550	-67	24	13
1,100	15	- 5	6
2,200	21	33	23
4,400	9	21	23
6¼ mi.	-83	-18	- 7

From this table it is seen: (i) that for the limit of 6¼ mi., better results would have been obtained by forecasting existing conditions for each period; (ii) that for 220 yd., existing conditions would have given better results at 12 hr. ahead; (iii) that for forecasts for 1 hr. ahead, the actual results show an improvement over persistence only for the limits 1,100 to 4,400 yd.; (iv) on the whole the forecasting was most successful in regard to the middle ranges 2,200 and 4,400 yd. and least successful for the extreme ranges. Similar data of visibility forecasts are available for a number of other stations, the checking of which can be carried out in the same manner.

3. Forecasts of elements subject to continuous variation.

An index of success for use with forecasts of scalar or vector elements was proposed by Priestley (1945). Let ε denote the standard error of the forecasts, i.e. the root mean square of the differences between the forecast and the actual values at time t; this being the period of the forecasts. Also let σ denote the root mean square of the variation between the values of the element initially and after time t. Then if there is no correlation between the forecast value and the forecast error, i.e. assuming the variations are not consistently under or over-estimated, we may express the variance of the element as

$$\sigma^2 = \varepsilon^2 + S^2,$$

where S^2 is the part which is successfully forecast. S may be termed the success, and its variance is given by

$$S^2 = \sigma^2 - \varepsilon^2$$

From this the index of success is written as

$$P = \frac{S}{\sigma} = \left\{1 - \frac{\varepsilon^2}{\sigma^2}\right\}^{\frac{1}{2}} \qquad \text{(9)}$$

Priestley discusses the effect of errors in the observations and also in the analyses on which the forecasts are based, and shows that disregard of the errors tends to make the formula underestimate the success. It is not however practicable to include the errors in a useful manner, owing to the uncertainty of those arising from analysis, so that the formula is best used in the form given although it will necessarily be subject to some limitation. It is seen that it is applicable to both scalars and vectors, provided that in the latter case ε and σ are taken to be the standard vector error and standard vector variation respectively.

The upper limit of the index is unity, and this is reached when the variance of the forecast errors becomes negligibly small in comparison with the time variance. Now suppose the forecast value of the element at time t is always the same as the initial value; then ε is by definition identical with σ. Thus for persistence forecasting, the value of the index is zero. On occasions a forecast will give a greater error than that which would arise from use of persistence. If this were to happen often enough, it would lead to ε becoming greater than σ, so that P^2 would become negative. This implies that the forecasting is worse than useless since the success is imaginary. In cases where forecasting is of value, the limits of the index are 0 and 1. As the index is non-dimensional and, as seen, measures the improvement over forecasting solely by persistence, it therefore constitutes an appropriate index for the measurement of success in forecasts of continuously varying elements.

3.1. Applications to forecasts of continuous elements.

3.1.1. Forecasts of height of pressure surface.

An evaluation of the index of success for forecasts of contour heights was made by Priestley (1945). Forecasts for 12 hour periods at Larkhill were made four times daily from prognostic contour ("prontour") charts, and the results compared with direct observations. Values of ε, σ and P were derived for each month from August to December, 1944, for pressures of 750,500 and 300 mb. These figures show no marked difference in the success between the three levels, but there is considerable variation of the index from month to month. This suggests that in order to derive a satisfactory value of the index, the average should be taken over a number of months, or over a number of years for the same month. The causes of the variations were discussed by Priestley, and it is clear that they are due to circumstances which cannot be eliminated by choice of index, but only by averaging over a longer period. Combined figures for the three months September - November, 1944, have been derived and are given in Table IX as an illustration.

Table IX - 12 hr. forecasts of height of pressure surface at Larkhill (September - November, 1944)

Pressure mb.	Standard deviation (feet)		Index of Success (P, per cent)
	Forecast error (ε)	12 hr. variation (σ)	
300	263	358	68
500	162	240	74
750	108	149	69

3.1.2. Forecasts of upper wind.

12 hour forecasts of upper wind at Larkhill were also dealt with by Priestley in a manner similar to that just described for the heights of pressure surfaces, and for the same periods. As with the height forecasts, there is no systematic variation of the index with height, while the variation from month to month is less marked in this case, although still considerable. It is clear that figures for individual months are influenced by the varying meteorological conditions and that longer periods are required to give reliable estimates. Figures for three months have been combined to give Table X.

/Table X.

Table X - 12 hr. forecasts of upper wind at Larkhill, September - November, 1944

Pressure mb.	Standard vector deviation		Index of Success (P, per cent)
	Forecast error (ϵ)	12 hour variation (σ)	
	knots	knots	
300	36	44	57
500	24	29	56
750	15	22	73

3.1.3. Forecasts of equivalent head winds.

An analysis of forecasts of equivalent head winds on the great circle route Shannon - Gander has been given by Harley (1952). From figures given for the standard deviation of the forecast errors and for the variation of equivalent head wind with time, Priestley's index of success may be computed. The results are shown in Table XI.

Table XI - Forecasts of equivalent head winds, Shannon - Gander (April, 1949 - May, 1951)

	Pressure	Winter	Spring	Summer	Autumn	Year
	mb.					
Standard error of forecasts (knots)	500	12.5	10.5	8.5	10.5	10.7
	700	8.7	7.8	6.5	8.1	7.9
S.D. of 24 hr. variation (knots)	500	14.8	13.5	9.6	12.3	12.5
	700	10.2	9.7	8.3	10.0	9.5
Index of success (per cent)	500	54	63	47	52	52
	700	52	59	62	59	56

In interpreting these figures, it is necessary to consider the method by which they were obtained. The forecast equivalent head winds were derived from composite contour charts, the time of flight being about 9 hours; the mid-time of these charts was about 24 hours ahead of the time of the basic charts from which they were prepared. On the other hand, the 'actual' equivalent head winds, with which the forecasts were compared, were derived from ordinary contour charts for a fixed time which agreed approximately with the mid-time of flight. The 24 hour variations were similarly dervied from the fixed time charts. The errors introduced by this use of fixed time charts were examined and found to be small. Further, all the winds used were geostrophic, but this approximation is held to be satisfactory when the results are averaged over a long route, since the ageostrophic components then tend to cancel out. It appears therefore that the results given in Table XI are reasonably accurate.

With the aid of known correlation coefficients (R_t) between equivalent head winds at times '0' and 't', Harley also derives values of the standard errors which would be expected if the forecasts were made statistically on the basis of a regression equation -

$$V_c' = R_t V_o \qquad \text{.....(10)}$$

/where

where v_o is the vector departure from the mean equivalent head wind at time '0' and v_t' is the expected departure at time 't'. In each season the standard errors so found exceed those associated with the conventional forecasts so that the index of success is reduced. It may be shown on the other hand, that the statistical forecasts necessarily improve on the persistence forecasts, i.e. those for which v_t' is assumed equal to v_o. Thus from the regression equation, the mean square error of the statistical forecasts, n in number, is:

$$\varepsilon^2 = \frac{1}{n}\sum(v_t' - v_t)^2 = \frac{1}{n}\sum(R_t v_o - v_t)^2$$

but since

$$\frac{1}{n}\sum v_o^2 = \frac{1}{n}\sum v_t^2$$

and

$$\sum v_o v_t = R_t \sum v_o^2$$

we find

$$\varepsilon^2 = \frac{1}{n}(1 - R_t^2)\sum v_o^2 .$$

For persistence forecasts, the mean square error is given by

$$\sigma^2 = \frac{1}{n}\sum(v_o - v_t)^2 = \frac{2}{n}(1 - R_t)\sum v_o^2$$

Therefore

$$\varepsilon^2 = \frac{1}{2}(1 + R_t)\sigma^2 \qquad \text{.....(11)}$$

so that ε is less than σ, since $R_t < 1$ for $t > 0$.
The extent of the improvement is shown by the values of the index of success which are given in Table XII, it being recalled that the index for persistence forecasts is zero. It will be noted also that each value of success in Table XII is less than the corresponding value in Table XI.

Table XII - Statistical forecasts of equivalent head winds

	Pressure	Winter	Spring	Summer	Autumn	Year
	mb.					
Standard error of forecasts (knots)	500	13.5	12.6	8.9	11.8	11.7
	700	9.3	9.2	7.6	9.4	8.9
Index of success (per cent)	500	41	36	37	28	35
	700	41	32	40	35	35

Note: The S.D. of the 24 hr. variation is the same as in Table XI.

/3.14.

3.14. Wind expectations for 36 hours.

At the present time, the equivalent head winds expected 36 hours ahead on various air routes over Europe are being supplied to airline operators on the basis of the regression equation (10). In Table XIII is given a statistical summary for two of the routes, namely Northolt to Barcelona with diversion to Nice, and Northolt to Copenhagen with diversion to Hamburg, for the period January to May, 1952.

Table XIII - Comparison of 36 hour wind expectations with actuals, for certain air routes at 10,000 feet.

ε = standard error of forecast winds in knots, σ = standard deviation of 36 hr. variation of wind, P = percentage index of success.

	Northolt - Barcelona - Nice			Northolt - Copenhagen - Hamburg		
1952	ε	σ	P	ε	σ	P
Jan.	9.7	11.2	50	12.6	14.6	51
Feb.	7.0	7.4	32	8.7	9.6	42
Mar.	7.5	8.9	54	7.2	7.5	28
Apr.	7.1	7.7	39	7.7	8.4	40
May	4.8	5.3	42	6.8	7.2	33
5 months	7.4	8.3	46	8.9	9.8	44

From 3.13 eqn. (11) we see that the index of success is given by

$$P^2 = 1 - \frac{\varepsilon^2}{\sigma^2} = \frac{1}{2}\left(1 - R_t\right) \qquad \text{(12)}$$

In the case under discussion the value of R_t is 0.60, and in consequence one would expect the value of P to be $(0.20)^{\frac{1}{2}}$ or 0.45. This agrees closely with the average value actually found over the five months, so confirming that the average success over a period of several months can be calculated in advance from the known value of the correlation coefficient of equivalent head winds with time.

Acknowledgements.

This paper was written at the suggestion of Mr. C.S. Durst, with whom the matter has been discussed. I also wish to thank Mr. G. A. Corby for providing data for forecasts of visibility at Northolt.

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At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 789 by Mr. P. Goldsmith. A copy is available in the Library of the Meteorological Office."

M.R.P. 789
S.C. III/147
25th March, 1953

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Some aircraft and surface meteorological observations made at Khartoum in July, 1952.

by
P. Goldsmith (Meteorological Office).

Summary.

An account is given of the meteorological measurements made at Khartoum in July, 1952 by a member of the Meteorological Research Flight in connection with tropical tests made by M.E. Dept. R.A.E. on an Avro-Ashton aircraft.

A description is given in Part I of aircraft measurements of temperature and humidity which were made between 20,000 and 39,000 feet on three occasions.

In Part II surface observations of solar radiation made with a Michelson actinometer are listed and the magnitude of the various radiation terms discussed.

Introduction.

During July, 1952, an Ashton, a four-engined experimental jet aircraft, was flown to Wadi-Seidna airfield near Khartoum from R.A.E. Farnborough for tropical trials. The aircraft carried a number of the staff of M.E. Dept. R.A.E. to carry out these trials which were concerned with the aircraft's skin temperatures and cabin air conditioning both in flight and on the ground.

A member of the Meteorological Research Flight, Mr. P. Goldsmith, accompanied the party to make temperature and humidity measurements in flight and observations of various meteorological variables including solar radiation on the ground. The measurements made are described and discussed below.

Part I.

Aircraft measurements of temperature and humidity at high altitudes.

The programme included three high level ascents in connection with the skin temperature work to be followed later by other ascents with the primary purpose of investigating the meteorological factors involved. Unfortunately a pressurisation fault limited this work to the first three flights. Readings of temperature with a Meteorological Office flat plate thermometer and of frost point with a pressurised Dobson-Brewer hygrometer were obtained on these flights and are listed in Table I.

/TABLE I.

TABLE I.

Flight No:-	1		2		3	
Date:-	12.7.52		14.7.52		15.7.52	
Time:-	1033 to 1225 Z		0621 to 0820 Z		1838 to 2034 Z	
HEIGHT feet ICAN.	Temp. °F	Frost Point °F	Temp. °F	Frost Point. °F	Temp. °F	Frost Point °F
39,000			-60.8	-83		
38,600	-57.2	-86				
38,000	-54.5	-90	-58.5	-67		
37,000	-51.8	-76	-53.0	-68		
36,500					-48.6	-62
36,000	-46.8	-72	-50.0	-71		
35,500					-43.8	
35,000	-41.9	-72	-44.2	-71	-41.7	-69
34,000	-37.8	-72	-39.3	-66	-35.4	-56
33,000	-33.4	-69	-34.8	-59	-32.8	-50
32,000	-29.6	-67.5	-29-8	-61	-27.8	-54
31,000	-25.1	-63 to-56	-26		-23.9	-53
30,000	-21.7	-58			-19.6	-50
28,000	-12.4	-63			-10.7	-40
27,000			- 8.4			
26,000	- 4.5	-56 to-53			- 5.3	-15
25,500			- 1.6			
24,000	- 0.7	-21			+ 3.4	-19
23,000			+ 5.1		+ 6.2	
22,000	+ 6.5		+ 7.4		+ 9.1	
21,700					+11.6	
21,000					+ 9.6	
20,000	+16.1		+13.5		+14.0	
19,000					+17.7	
18,000			+22.0		+21.2	
17,000					+25.8	
16,000			+30.2		+29.6	
15,000	+33.6		+35.1		+34.7	
14,000	+37.6		+39.8		+39.3	
13,000	+42.4				+42.4	
12,000	+47.2		+49.6		+46.6	
11,000	+51.9				+51.8	
10,000	+54.3		+59.1		+57.1	
9,000	+58.9		+63.5		+63.2	
8,000	+63.9		+67.5			
7,000	+70.1		+71.0		+73.8	
6,000	+76.1				+79.4	
5,000	+80.5		+80.5			
4,500			+82.8			
4,000	+87.0				+89.3	
3,000			+87.3			
Surface	+105.0					

/Three

Three tephigrams are given (Figs.1, 2 and 3) on which these readings are plotted together with other details of the flights. For comparison purposes the results of the nearest radio-sonde ascents have also been plotted on these diagrams (Nairobi is about 1,000 miles SSE of Khartoum and Aden about 800 miles ESE of Khartoum).

Discussion of observations.

Flight No.I. 12/7/52. (1233 to 1425 local time)

The weather was fine with good visibility and only 2-3/8 cirrus cloud. Shade temperatures at the surface were between 102 and 106°F and there was a light northwesterly wind.

Readings of temperature and airspeed were taken during the climb up to 20,000 feet.

Above 20,000 feet the Dobson-Brewer hygrometer was prepared for use whilst the aircraft ascended to 38,000 feet and the M.E. Dept. staff made their non-meteorological observations. Readings of temperature and frost point were taken on the descent from 38,000 feet to 20,000 feet.

Considering the observations in conjunction with the Aden 0300Z and 1500Z ascents and the Nairobi 0300Z ascent the following comments are given.

(1) The lapse rate above 20,000 feet was mainly between the wet adiabatic and dry adiabatic values and there was a layer of very much reduced lapse rate from 24,000 feet to 26,000 feet which suggests a change of air mass at that level.

(2) There is good agreement between the aircraft sounding and the Aden ascents, particularly the 0300Z ascent at all levels apart from the region where the effect of the surface is predominant. The apparent change of air mass occurs somewhere lower at Aden. It was better marked at 0300Z than 1500Z, and corresponds with a change of wind direction from north east to east with a decrease of speed in the boundary region.

(3) Above 26,000 feet there is fair agreement also with the Nairobi ascent suggesting that, at these levels, the air at all three stations is of similar origin. The agreement in temperatures may however be fortuitous as at that level the Nairobi wind has a marked southerly component. Above about 32,000 feet there is a tendency for the Nairobi temperatures to be less than the others. Wind speeds there decrease with height while a strong easterly current increases with height at Aden.

(4) Considering now the humidity readings, since the tropopause on this occasion was estimated from the radio-sonde data to be about 57,000 feet, these readings can be regarded as referring to the middle troposphere only. The lowest frost point values of -80°F to -90°F which were found at 37,000 feet to 38,000 feet are normally only found at similar heights in England i.e. near the tropopause level. The rapid decrease of humidity with height at 24,000 feet is also evident in the 0300Z Aden ascent which again suggests a change in air mass at that level.

Flight No.II. 14/7/52. (0821 to 1020 local time)

The weather was fine with good visibility with only 2-3/8 cirrocumulus and altocumulus. The surface temperature increased from 95°F to 100°F and the wind was calm.

The flight procedure was similar to that of Flight No.I with temperature readings on the ascent to 27,000 feet, followed by a period of non-meteorological readings and finally temperature and frost point readings on the descent from 39,000 feet to 31,000 feet. No readings were taken between 31,000 feet and 27,000 feet owing to a failure of the hygrometer.

/Considering

Considering the observations in conjunction with the nearest Nairobi and Aden ascents the following comments are given.

(1) From the surface to about 20,000 feet there was a steady lapse rate and then from this level to about 25,000 feet there was a considerable decrease in lapse rate. From 26,000 feet to 39,000 feet the lapse rate was nearly wet adiabatic.

(2) The general form of the temperature sounding is very similar to the Aden ascent except that in the latter there was an inversion from about 14,000 feet to 15,000 feet accompanied by a change from a northeasterly to an easterly wind regime. The Khartoum and Aden temperature soundings show good agreement from about 8,000 feet to 14,000 feet and again above 26,000 feet .

(3) Above 26,000 feet there is also good agreement with the Nairobi ascent, suggesting that at lower levels Aden and Khartoum were in air of the same origin and at higher levels both they and Nairobi were in air of a different origin from that at lower levels.

(4) The humidity readings present no points of special interest except that, apart from the highest reading, there is a tendency for the relative humidity to increase towards 39,000 feet and some cirrus was estimated at 40,000 feet or above.

Flight No.III. 15/7/52. (2038 to 2234 local time)

This was a night flight. The weather was again fine with only 2-3/8 cirrostratus and altocumulus. This latter cloud was found to be between 20,000 feet and 22,000 feet above the ground.

Temperature readings were taken on the ascent and temperature and frost point readings at the higher levels of the descent, the procedure being very similar to that of the two previous flights. This flight had to be terminated before completion of the observations owing to a fault developing in the aircraft's pressurisation system.

The ascent showed the following features.

(1) Apart from the lowest layers there was a dry adiabatic lapse rate to about 12,000 feet followed by a steadily decreasing lapse rate to 21,000 feet where there was an inversion. Above about 23,000 feet there was a wet adiabatic lapse rate with good agreement between the Khartoum aircraft measurements and both the Nairobi and Aden soundings.

(2) The inversion layer at 21,000 feet probably coincided roughly with the tops of the altocumulus cloud. At Aden there was an inversion at about 19,000 feet with a northeasterly wind regime below and an easterly wind regime above. This feature was observed on all three flights and again suggests air of different origin above and below 20,000 feet to 25,000 feet at both Aden and Khartoum.

(3) In the Khartoum sounding there was a large lapse of humidity at 26,000 feet i.e. just above the transition layer. This effect is not noticeable on the Aden ascent although a similar large decrease of humidity was observed above an inversion at 16,000 feet on the Nairobi ascent.

/Discussion.

Discussion.

(1) It has been shown in M.R.P. 729 (Ref.1) that easterly winds are persistent at Aden in July up to great heights. The vector resultants quoted in Table 1 of that paper for July are as follows

Pressure in mb.	850	700	500	300	200	150	100
Vector resultant direction.	278°	29°	45°	80°	90°	93°	94°
Speed(kt.)	5	3	13	23	48	71	66

From this table it is apparent that the onset of the main easterly stream occurs between 500 and 300 mb. i.e. between 20,000 and 30,000 feet. The Aden soundings discussed above all exhibited this feature and the evidence is that the addendum by C.S. Durst to M.R.P. 729 viz., that "between 35,000 and 40,000 feet easterly winds of the same magnitude as at Aden extend over ..... the Sudan", can probably be extended to include the range 25,000 to 35,000 feet in July.

(2) It has been demonstrated that providing the necessary supplies of liquid oxygen can be obtained locally the Dobson-Brewer hygrometer can readily be operated in the tropics. Provided that attention is given to the effect of the high temperatures (circa + 130°F) and humidities in the aircraft on the ground on components such as rubber seals, wiring etc. there should be no difficulty in extending this work to tropical regions if this should be required. A possible limitation on the use of the instrument is discussed below.

(3) It can be seen from Figs.1, 2 and 3 that frost points of about -80°F have been obtained at 38,000 feet in Khartoum with an air temperature of -60°F. The tropopause was at about 57,000 feet with an air temperature of -110°F approximately. It seems that unless the tropopause region in the tropics is always saturated frost points of -120°F or lower would be expected there. Although considerable cirrus is found in the tropics, it is not considered likely that saturation will always exist near the tropopause there.

It is interesting incidentally to note that as temperatures of -130°F have been reported at moderate altitudes during the polar night (Ref.2) this instrument may also fail at high latitudes, and for accurate humidity readings at high altitudes in both tropical and polar regions there will be a requirement for a new type of instrument.

Part II

Measurements on the ground.

Introduction.

In conjunction with the aircraft skin temperature trials on the ground which were made by the M.E. Dept. R.A.E. from July 12th to July 17th, 1952 meteorological measurements were made to obtain information on the radiation field in which the aircraft was situated. Their application to the aircraft problem is outside the scope of this report, but the measurements themselves may be of general interest.

/The

The components of radiation to be considered were

1. Short wave radiation (0.3μ to 3.0μ wavelength approximately)

   (a) The direct solar beam.

   (b) Scattered solar radiation.

   (c) Solar radiation reflected from the earth's surface.

2. Long wave radiation (wavelengths greater than 3.0μ .)

   (a) Atmospheric long wave radiation.

   (b) Long wave radiation from the ground.

The measurements and the estimation of the magnitudes of each of the components which were not measured directly are described below, and Fig. 4 shows their computed values throughout the day at Khartoum on 13th and 14th July, 1952. For these estimations and advice on this part of the work I am indebted to Dr. Robinson, the Superintendent of Kew Observatory and Mr. J MacDowall also of Kew Observatory who made the detailed computations.

1 (a) The direct solar beam.

A Michelson bimetallic actinometer, which had previously been used in Khartoum for other work, was borrowed from Kew Observatory and used for regular measurements of the intensity of the direct solar beam. The sun's zenith angle was also noted at each reading. This instrument was calibrated at Kew before and after use in Khartoum by comparison with a sub-standard Angstrom pyrheliometer. The readings obtained are listed in Table II.

1 (b) Scattered solar radiation.

This was estimated using an empirical relation between the intensity of the direct solar beam and the intensity of diffuse radiation given in Ref. 4.

1 (c) Short-wave radiation reflected from the ground.

It was assumed that the sand surface would have a reflectivity of 0.25 (Ref. 5) i.e. that this component was 0.25 times the sum of the direct solar beam resolved on to a horizontal surface and the diffuse radiation.

2 (a) Atmospheric long-wave radiation.

This component which is dependent mainly on the absorption bands of water vapour and carbon dioxide in the atmosphere is readily estimated by means of a radiation chart when an upper air-sounding is available. In view of the general good agreement between the upper air measurements of Flights I, II and III and the relevant Aden radio-sonde ascents it was considered that a reasonably accurate value (say ± 15%) of this component could be obtained by using the Kew Radiation Chart in conjunction with estimated values of upper air temperatures and humidities based on Khartoum surface measurements and Aden radio-sonde measurements on these occasions.

2 (b) Long-wave radiation from the ground.

Ground temperatures were measured at regular intervals by means of a thermometer with its bulb immersed to a depth of one inch in the sand. This component was then calculated from the sand temperatures on the assumption that for the wavelengths involved the sand radiates as a black body.

/Results.

Results.

Using these computed values approximate values of the various radiation components at Khartoum in July are compared with corresponding mean values for cloudless days at Kew in m.watts/cm ^{2}.

		Khartoum.	Kew.
(a)	Direct solar beam (local noon)	90	80
(b)	Scattered solar radiation	15	15
(c)	Reflected short-wave radiation	25	15
(d)	Atmospheric long-wave radiation	40	30
(e)	Ground long-wave radiation	70	55

It should be emphasised that these values for Khartoum are only approximate although sufficiently accurate for most purposes. It is recommended that if future work of this type is undertaken a direct measurement of the diffuse sky radiation be obtained. In addition measurements of the reflectivity of the soil would be an advantage.

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No. 5. Smithsonian Met. Tables, 1951, after T.H. Macdonald.

Acknowledgements.

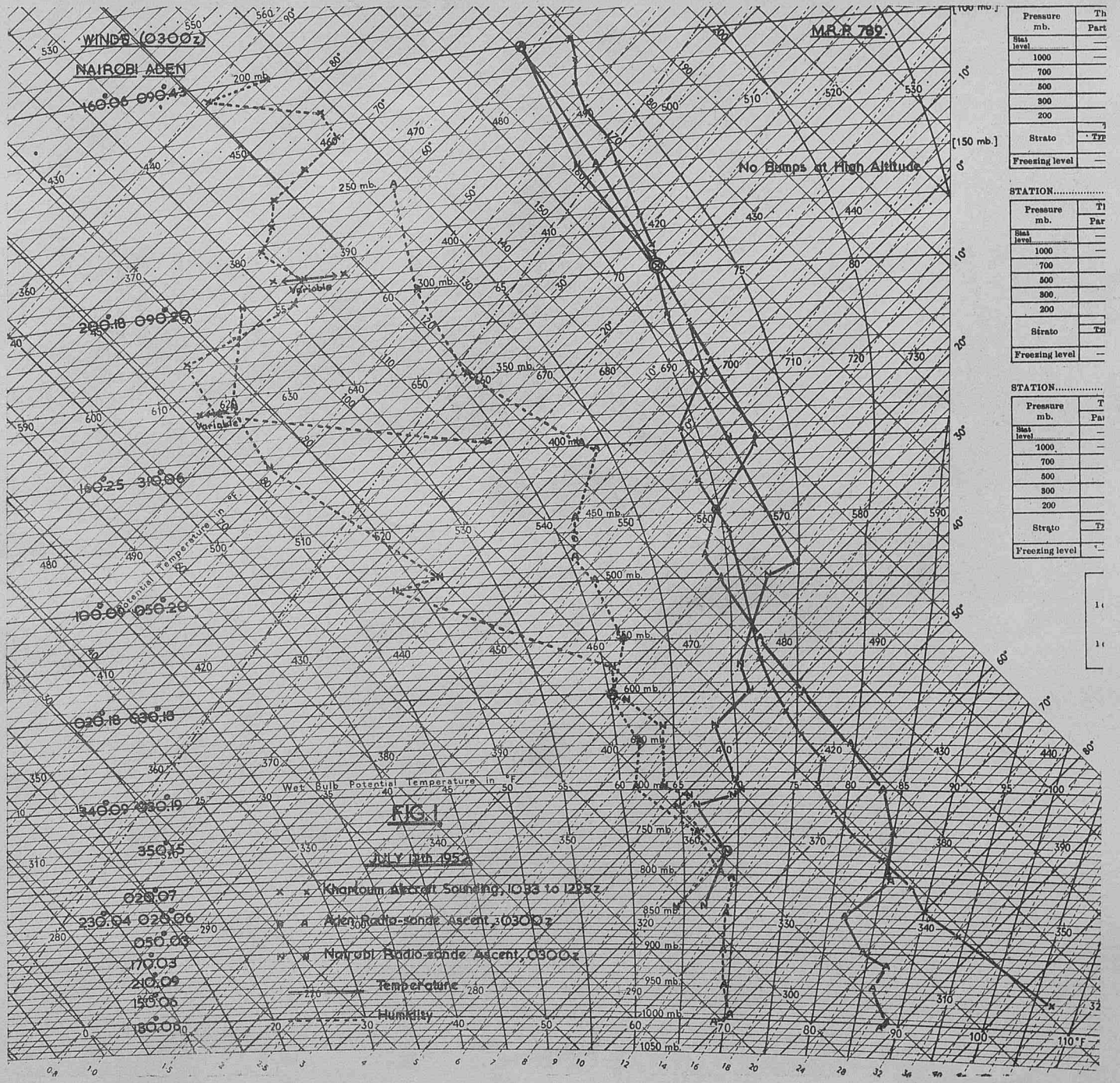
I wish to express my appreciation to Mr. R.J. Murgatroyd, S.Met.O of the Met. Research Flight, for arranging for me to accompany the Ashton to Khartoum and for his advice in preparing this paper.

/TABLE II.

Actinometer Results. TABLE II.

July 12th, 1952.

Local Time.	Reading mw/cm^2	Shade Temp. °F	Sand Temp. °F	Elevation Angle Degrees (from zenith)	Sky.
1000	82.5	98	124	27	Clear
1100	85.8	99	128	15	"
1130	84.7	102	133	09	"
1204	85.8	102	135	06	"
1450	74.8	109	136	42	"
1520	61.6	107	136	49	"
1600	57.2	106	131	58	"
1620	50.6	108	130	62	"
July 13th, 1952.					
1030	83.6	99	125	21	Clear)
1045	83.6	100	133	18	Trace Ci)
1100	88.0	102	134	14	")
1115	86.9	102	135	12	4/8 Cc)
1130	85.8	103	136	09	3-4/8 Cc)
1145	88.0	103	137	06	2/8 ")
1200	88.0	106	138	07	2/8 ") Sun
1215	89.1	105	135	09	3/8 ") clear.
1230	88.0	104	139	12	3/8 ")
1240	85.8	106	139	14	3/8 ")
1300	84.7	106	141	18	3/8 ")
1315	33.0	107	140	20	5/8 Ac)
1330	4.4	106	136	24	6/8 ") Sun
1345	2.2.	106	132	27	7/8 ") obscured
1400	8.8	105	130	30	7/8 ")
1410	44.0	-		33	7/8 ")
1415	81.4	107	132	34	7/8 ")
1430	61.6	108	136	38	7/8 ") Sun
1445	77.0	106	133	41	7/8 ") Clear.
1500	80.3	107	132	45	7/8 "
July 14th, 1952.					
1055	90.2	101	129	16	Trace Ci
1107	90.2	101	130	13	Clear
1118	90.2	101	131	10	"
1134	92.4	102	132	08	"
1140	90.2	104	135	06	"
1200	92.4	106	136	06	"
1220	92.4	106	137	08	"
1230	90.2	106	139	10	"
1245	88.0	104	139	14	"
1300	88.0	105	139	17	"
1315	89.1	106	139	20	"
1330	85.8	108	139	23	"
1345	85.8	108	139	28	"
1400	83.6	109	138	31	"
1415	83.6	108	138	34	")Tiny Cu
1430	83.6	108	137	38	")forming
1445	81.4	109	135	41	") on
1500	78.1	108	135	44	")horizon.
1515	77.0	108	134	47	
17th July, 1952.					
1420	77.0	101	121	35	Trace Ac
1440	22.0 to 66.0			40	)
1500	37.4 to 52.8	101	120	44	) Sun obscured
1515	71.5	103	121	47	) by occasional
1530	17.6 to 52.8	101	119	51	) Ac



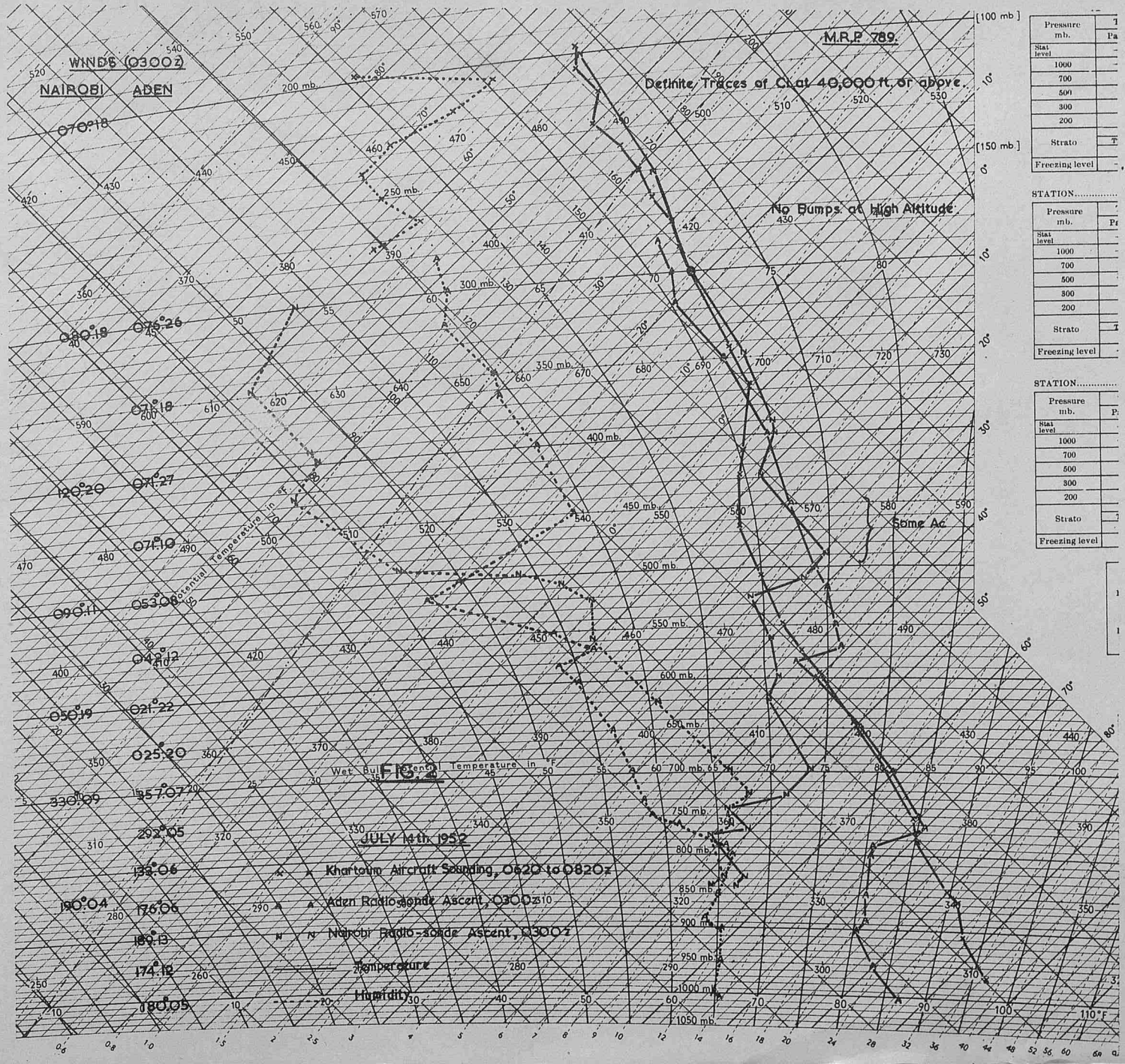
Pressure mb.	Th / Part
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Tro
Freezing level	—

STATION.....

Pressure mb.	Th / Par
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Tro
Freezing level	—

STATION.....

Pressure mb.	T / Pa
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Tr
Freezing level	—



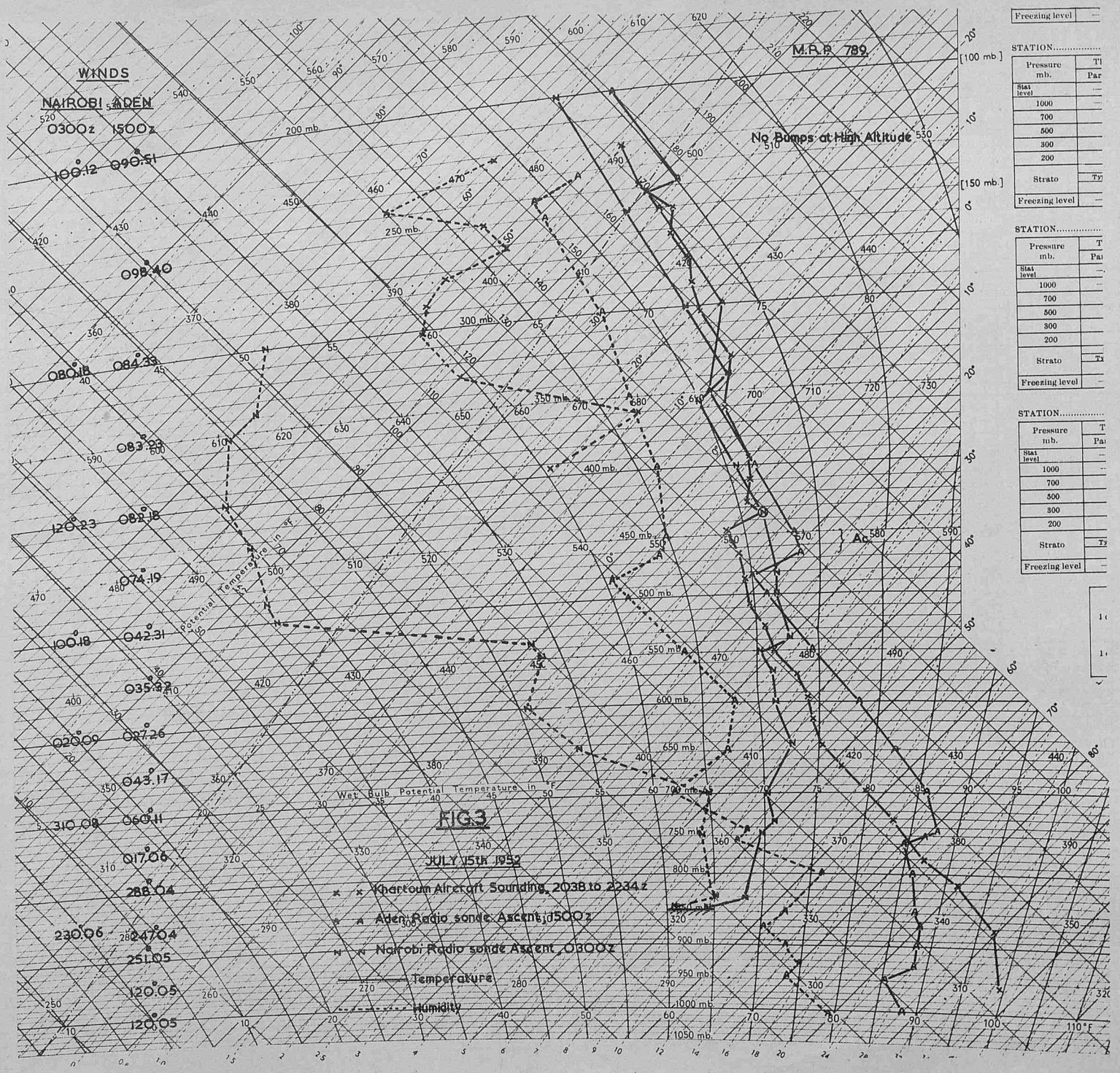
Pressure mb.	T / Pa
Stat level	
1000	
700	
500	
300	
200	
Strato	T
Freezing level	

STATION.....

Pressure mb.	T / Pa
Stat level	
1000	
700	
500	
300	
200	
Strato	T
Freezing level	

STATION.....

Pressure mb.	T / Pa
Stat level	
1000	
700	
500	
300	
200	
Strato	
Freezing level	



Freezing level	—

STATION.....

Pressure mb.	T / Par
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Ty
Freezing level	—

STATION.....

Pressure mb.	T / Pa
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Ty
Freezing level	—

STATION.....

Pressure mb.	T / Pa
Stat level	—
1000	—
700	
500	
300	
200	
Strato	Ty
Freezing level	—

M.R.P. 789.

A Reconstruction of the Radiation Field at Khartoum on the 13th and 14th July 1952.

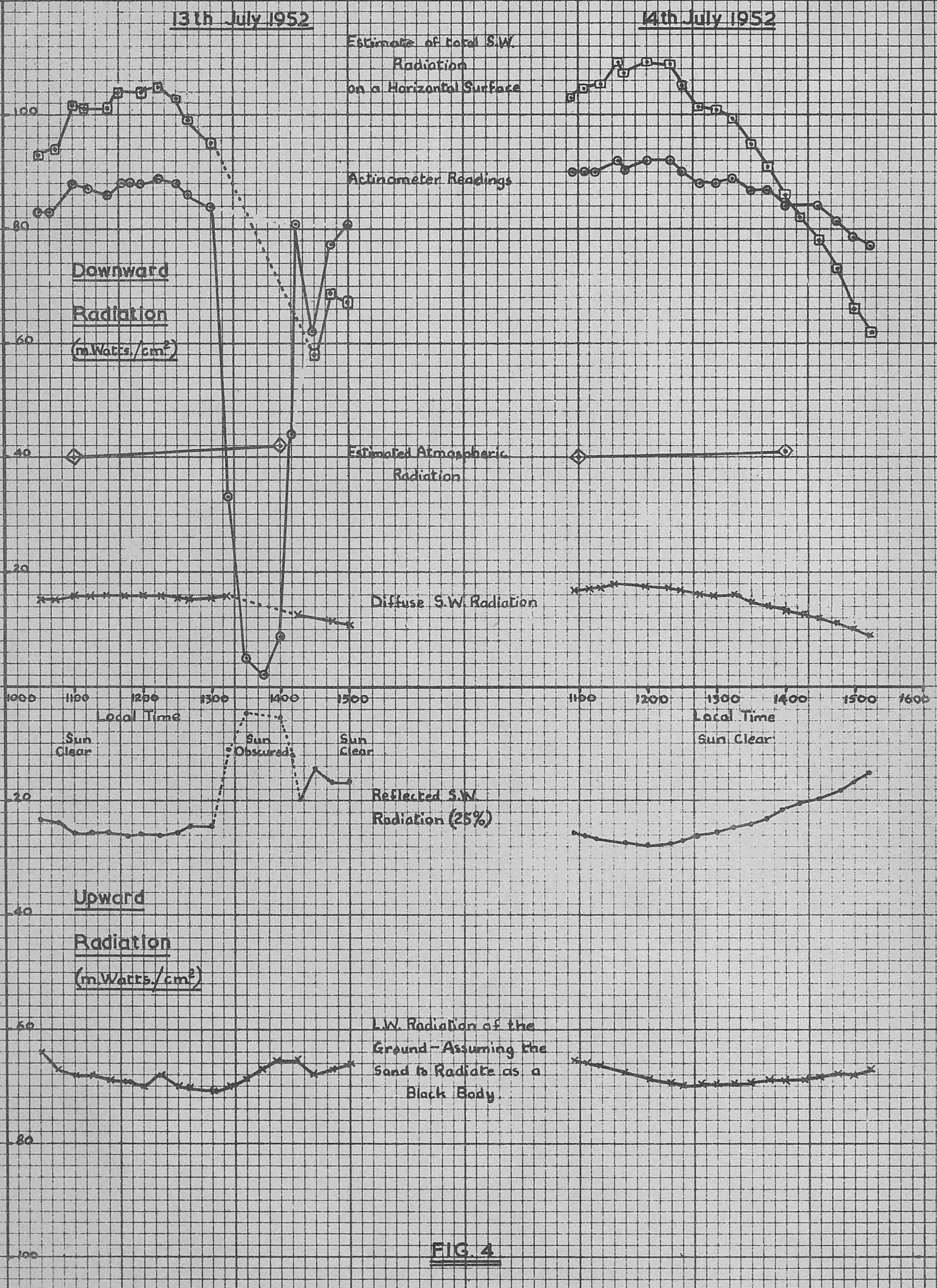


FIG. 4

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 790 by Mr. F.H. Bushby & Miss Mavis K. Hinds. A copy is available in the Library of the Meteorological Office."

M.R.P. 790
S.C.II/139
24th March, 1953.

AD 44249

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Computation of the 500 mb height tendency in a baroclinic atmosphere, using an electronic computer

by
F.H. Bushby and Mavis K. Hinds (Meteorological Office)

1. Introduction

In a recent paper Sawyer and Bushby (1) have derived, subject to various approximations, a set of simultaneous partial differential equations for forecasting the contour height of an isobaric surface representative of the mid-troposphere and the thickness of the layer from 1000 mb to that isobaric surface. The method is one in which the instantaneous rates of change of the contour height and thickness pattern are first computed at time t, and from these, new contour and thickness patterns are computed for time $t+\delta t$. The process is then repeated n times, and forecast charts are produced for time $t+n\delta t$.

The construction of a programme for the solution of this set of equations by an electronic computing machine is a formidable task. It was therefore decided to prepare one programme for the solution of the equation for the rate of change of contour height and another for the solution of the equation for the rate of change of thickness before proceeding with the step by step process of time integration. The present report describes the computation of the 500 mb height tendency on LEO for three synoptic situations, and the computed results are compared with the actual changes that occurred. The effect of assuming wrong boundary conditions is discussed, as is the effect of neglecting the vertical advection of vorticity.

2. The Method

In the model adopted by Sawyer and Bushby (1), the state of the atmosphere at a given instant is specified by three parameters; viz. h_m, the height of the central pressure of the troposphere; h', the thickness of the bottom half of the troposphere; and a stability parameter. In the present work h_m has been taken as the height of the 500 mb contour surface, and h' as the thickness of the layer from 1000 mb to 500 mb. It would probably be more accurate to use the 600 mb contour chart and the 1000 - 600 mb thickness chart but these were not available.

Conformally transforming the northern hemisphere by a stereographic projection from the south pole on to a plane perpendicular to the earth's axis and passing through the north pole, the equation for $\frac{\partial h_m}{\partial t}$ is given

/by

by

$$\nabla^2\frac{\partial h_m}{\partial t} + J\left(h_m, \beta^2 g l^{-1}\nabla^2 h_m + l\right) + \tfrac{1}{3}J\left(h', \beta^2 g l^{-1}\nabla^2 h'\right) - \frac{8\varpi_m \nabla^2 h'}{3(p_0 - p_1)} = 0 \qquad (1)$$

where ∇^2 is the two-dimensional Laplacian operator referred to rectangular Cartesian co-ordinates; J is the Jacobian operator; β is the magnification factor $\sec^2\left(\frac{\pi}{4} - \frac{\theta}{2}\right)$, θ being the geographical latitude; g is the acceleration due to gravity; l is the Coriolis parameter; ϖ_m is the value of $\frac{dp}{dt}$ at the level h_m; and $p_0 - p_1$ is the difference of pressure through the troposphere, in this case taken as 800 mb.

Preliminary computations showed that the last term on the left-hand side of Equation 1, which represents the vertical advection of vorticity, was probably small compared with the remaining terms, and it was therefore initially omitted from the calculations. This in fact simplified the calculations a great deal, for to compute ϖ_m it would have been necessary first to calculate $\frac{\partial h'}{\partial t}$. This simplification also made the use of the stability parameter unnecessary.

A programme, MET 4, was therefore made to solve

$$\nabla^2\frac{\partial h_m}{\partial t} + J\left(h_m, \beta^2 g l^{-1}\nabla^2 h_m + l\right) + \tfrac{1}{3}J\left(h', \beta^2 g l^{-1}\nabla^2 h'\right) = 0 \qquad (2)$$

at a network of 12 x 8 points. Equation 2, with the last term omitted, is the Charney (2) equation for an equivalent barotropic atmosphere. The grid length used, a, was about 160 miles in middle latitudes. Use was made of preset parameters so that the programme can easily be adapted for a different number of grid points, the only limitation being the storage capacity, 2048 locations, of the computing machine. The programme occupied 480 storage locations whilst the data, intermediate and final results for an r x s grid would occupy (6 rs + 16 r + 24 s + 52) locations, 1012 in this case.

The Jacobians were printed as intermediate results so that their relative magnitudes could be compared. The Poisson type differential equation was solved by a slight modification of the iterative process due to Liebmann (3). The computed values of $\frac{\partial h_m}{\partial t}$ were printed after 30, 40 and 50 iterations. The greatest changes in $\frac{\partial h_m}{\partial t}$ between 30 and 40 iterations were about 10%, but there were no appreciable changes between 40 and 50 iterations. If no intermediate results were printed and the values of $\frac{\partial h_m}{\partial t}$ were only printed after 40 iterations, the programme would take about 12 minutes to run.

In order to solve Equation 2 it is necessary to assume the values of $\frac{\partial h_m}{\partial t}$ along the boundary. In the present computations $\frac{\partial h_m}{\partial t}$ was taken to be zero at the 44 points exterior to the 12 x 8 grid at which values were computed. In fact it was known that $\frac{\partial h_m}{\partial t}$ was not zero on the boundary and in order to estimate the effect of using wrong boundary conditions, the true values of $\frac{\partial h_m}{\partial t}$ on the boundary were estimated from working charts, and the necessary corrections to the computed values of $\frac{\partial h_m}{\partial t}$ were calculated by hand. The effect of these corrections is discussed in section 3.

/A

A programme for computing the thickness tendency on the lines suggested by Sawyer and Bushby (1) has subsequently been made, and the machine was instructed to compute the term $\left(-\frac{8\varpi_m a^2\nabla^2 h'}{3(p_0-p_1)}\right)$. This has now been done for the situations described in this report. The computation of the thickness tendencies will be described in a later report, but the values of $\left(-\frac{8\varpi_m a^2\nabla^2 h'}{3(p_0-p_1)}\right)$ at grid points are given in the appendix, together with corresponding values of the two Jacobians of Equation 2. The corrections to the computed values of $\frac{\partial h_m}{\partial t}$ necessary because of the neglect of this term have been worked out by using hand relaxation, and thus solutions to Equation 1 were obtained. The values of $\frac{\partial h_m}{\partial t}$ obtained from Equation 1 have been compared with those obtained from Equation 2 and the results are shown in Tables I and II.

3. Results

The values of the three vorticity advection terms of Equation 1 for the three situations considered are tabulated in the appendix. In each case the term $a^2 J(h_m, \beta^2 g l^{-1}\nabla^2 h_m + l)$, which represents the horizontal advection of the 500 mb absolute vorticity field, is the most important. The term $\frac{a^2}{3}J(h', \beta^2 g l^{-1}\nabla^2 h')$ adds a contribution at about 60% of the grid points, whilst the term $-\frac{8\varpi_m a^2\nabla^2 h'}{3(p_0-p_1)}$, representing the vertical advection of vorticity, makes a contribution at approximately 30% of the grid points. Either of these terms can, in certain limited areas, be numerically greater than the first Jacobian.

The values of the 500 mb height tendency computed from solving Equation 2 assuming all the boundary values are zero are shown in Figures 1c, 2c and 3c. The synoptic situations to which they refer are shown in Figures 1a, 2a and 3a and the actual 500 mb height tendency fields, averaged over 12 hour periods centred at the times for which the computations were made, are shown in Figures 1b, 2b and 3b. It can be seen from the diagrams that there is a good correspondence between the computed and actual tendency fields in all three cases. At first, it may seem inconsistent that an equation in which the main terms are those of Charney's equation for a barotropic model atmosphere gives much better results than those found by Charney (2) or Bushby (4) when solving the barotropic equation. The additional terms are obviously not responsible for all the improvement, and it therefore seems that the use of a much smaller grid is an important factor. The 450 mile grid length used in previous work by Charney and Bushby appears to be much too coarse to represent adequately the vorticity field by finite difference approximations, and Charney (5) has also come to a similar conclusion.

Correlation coefficients have been calculated comparing the computed tendency with the actual tendency, and regression coefficients were calculated for the best fitting straight line giving the actual change (Y feet per hour) in terms of the computed change (X feet per hour) in the form

$$Y = bX + c \qquad (3)$$

and the results are shown in Table I. The R.M.S. differences between the actual and computed tendencies are given in Table II. The notation used in Tables I and II is as follows:-

A ≡ Actual $\frac{\partial h_m}{\partial t}$ (ft./hr.) obtained from change in h_m over a 12 hour interval.

B ≡ Computed $\frac{\partial h_m}{\partial t}$ (ft./hr.) using Equation 2, zero boundary values.

C ≡ Computed " (ft./hr.) using Equation 1, zero boundary values.

D ≡ Computed " (ft./hr.) using Equation 2, actual boundary values.

E ≡ Computed " (ft./hr.) using Equation 1, actual boundary values.

/Table 1

Table I

Correlation and regression coefficients of actual and computed 500 mb height tendencies

		1500 G.M.T. 14.3.49			1500 G.M.T. 27.1.52			1500 G.M.T. 20.12.51		
Y	X	r	b	c	r	b	c	r	b	c
A	B	0.67	0.85	3.1	0.83	1.05	5.6	0.70	0.76	4.3
A	C	0.64	0.72	6.2	0.78	0.99	9.8	0.69	0.78	7.2
A	D	0.84	0.80	1.4	0.83	0.85	2.2	0.77	0.74	4.3
A	E	0.82	0.74	4.1	0.80	0.85	6.2	0.77	0.76	7.1

r = correlation coefficient

Table II

Root mean squares in ft./hr.

	1500 GMT 14. 3.49	1500 GMT 27. 1.52	1500 GMT 20.12.51
A	21.3	15.8	26.5
A - B	14.5	10.3	16.6
A - C	16.3	14.0	17.5
A - D	11.1	9.5	15.4
A - E	12.5	12.2	16.1

From an inspection of Tables I and II it is evident that the solutions of Equation 2 for $\frac{\partial \zeta_m}{\partial t}$ assuming zero boundary conditions agree quite well with the actual values. The incorporation of the term representing the vertical advection of vorticity does not significantly affect the correlation coefficients but this term has increased the R.M.S. difference between the actual and computed tendency, although in two of the cases the increase was only slight. However, as one is comparing a computed instantaneous tendency with an actual tendency averaged over a 12 hour period, it does not follow that the magnitude of the computed tendency should agree with the measured value. Thus although the instantaneous tendencies are apparently made worse by the incorporation of the vertical advection of vorticity, it does not necessarily follow that they are in fact worse. This can only be decided after time integration has been carried out.

Nevertheless it seems quite obvious from the available information that the assumption of the correct boundary conditions does significantly increase the agreement between the computed and actual tendency fields, and it will be necessary to take some notice of this when attempting time integration.

4. Conclusions

The vertical advection of vorticity seems to be insignificant in two of the three situations; in fact its inclusion slightly lessens the degree of correspondence between the actual and computed tendencies. However, this may not always be so. The effect of boundary conditions seems important if a small area is considered, and before time integration is started it would seem necessary either to increase the area under consideration so that the effect of boundary conditions would not affect the central area, or to make a preliminary estimate of expected changes around the boundary and feed this into the machine.

/As

As the time factor involved in computing 500 mb height tendencies at 96 grid points is about 12 minutes machine time, it is quite practicable for 500 mb height tendency charts to be produced on a routine basis in time for them to be of use to the forecaster. Obviously they would not be of as much use as prognostic charts for 24 hours ahead, but if the degree of success achieved with the situations so far examined could be maintained, they might provide a useful tool for the forecaster.

Nevertheless, the most important conclusion from the present work is that it shows that the model gives a fair representation of the 500 mb tendencies in the situations to which it has been applied, and justifies continuing the experiment. The next step, the computation of thickness tendencies, is well advanced and will be reported in due course; integration with respect to time will then be undertaken if the results continue to be encouraging.

/REFERENCES

-1	2	-1	0	1	0	-2	1	1	0	1	0
-1	0	0	0	0	0	0	0	0	-1	0	0
2	0	0	1	0	0	1	1	-1	0	0	0

Table IIIC

$$-\frac{8a^2\varpi_m\nabla^2 h'}{3(p_0-p_1)}$$

in ft/hr for 1500 G.M.T. 14.3.1949

-20	-1	-2	5	1	-25	9	1	3	-2	-1	-3
-4	-8	0	1	4	2	-10	1	1	0	-1	-1
-2	-4	1	-1	-2	0	-4	2	2	0	-2	-1
-3	0	1	7	6	1	-1	0	-20	-1	3	1
0	0	1	0	0	1	-3	-17	-4	0	0	0
-1	0	0	0	0	1	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0

/Table IVA

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From an inspection of Tables I and II it is evident that the solutions of Equation 2 for $\frac{\partial \zeta_m}{\partial t}$ assuming zero boundary conditions agree quite well with the actual values. The incorporation of the term representing the vertical advection of vorticity does not significantly affect the correlation coefficients but this term has increased the R.M.S. difference between the actual and computed tendency, although in two of the cases the increase was only slight. However, as one is comparing a computed instantaneous tendency with an actual tendency averaged over a 12 hour period, it does not follow that the magnitude of the computed tendency should agree with the measured value. Thus although the instantaneous tendencies are apparently made worse by the incorporation of the vertical advection of vorticity, it does not necessarily follow that they are in fact worse. This can only be decided after time integration has been carried out.

Nevertheless it seems quite obvious from the available information that the assumption of the correct boundary conditions does significantly increase the agreement between the computed and actual tendency fields, and it will be necessary to take some notice of this when attempting time integration.

4. Conclusions

The vertical advection of vorticity seems to be insignificant in two of the three situations; in fact its inclusion slightly lessens the degree of correspondence between the actual and computed tendencies. However, this may not always be so. The effect of boundary conditions seems important if a small area is considered, and before time integration is started it would seem necessary either to increase the area under consideration so that the effect of boundary conditions would not affect the central area, or to make a preliminary estimate of expected changes around the boundary and feed this into the machine.

/As

APPENDIX

The arrays of points in the following tables refer to the 12 x 8 grid points used in the computation. The exterior boundary of the area considered is shown in Figure 1c.

Table IIIA

$a^2 J(h_m, \beta^2 g l^{-1} \nabla^2 h_m + l)$ in ft/hr for 1500 G.M.T. 14.3.1949.

-2	-24	-10	24	69	-5	-21	6	24	14	16	17
-20	-5	3	14	-5	6	43	-40	-12	50	34	-5
-24	-28	11	14	4	25	11	-4	-23	3	7	0
6	10	-11	1	23	32	12	-73	-71	6	33	34
6	4	8	16	9	11	-26	-14	3	-3	10	-3
-6	-2	2	-3	5	-8	-20	8	-1	-1	0	-2
-3	-2	-4	-5	0	-3	-2	-3	6	5	1	2
7	3	-11	14	0	0	0	5	4	-2	-1	-2

Table IIIB

$\frac{1}{3} a^2 J(h', \beta^2 g l^{-1} \nabla^2 h')$ in ft/hr for 1500 G.M.T. 14.3.1949.

7	13	0	7	23	13	-25	-12	7	2	1	3
-2	-1	4	4	2	10	7	-16	-4	6	1	-1
-2	-3	4	2	-2	4	10	1	-6	4	6	-5
-1	1	-2	-2	8	7	0	-16	-5	3	0	6
1	0	2	4	-2	-5	-5	3	11	3	-3	-1
-1	2	-1	0	1	0	-2	1	1	0	1	0
-1	0	0	0	0	0	0	0	0	-1	0	0
2	0	0	1	0	0	1	1	-1	0	0	0

Table IIIC

$-\frac{8 a^2 \varpi_m \nabla^2 h'}{3(P_0 - P_1)}$ in ft/hr for 1500 G.M.T. 14.3.1949

-20	-1	-2	5	1	-25	9	1	3	-2	-1	-3
-4	-8	0	1	4	2	-10	1	1	0	-1	-1
-2	-4	1	-1	-2	0	-4	2	2	0	-2	-1
-3	0	1	7	6	1	-1	0	-20	-1	3	1
0	0	1	0	0	1	-3	-17	-4	0	0	0
-1	0	0	0	0	1	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0

/Table IVA

Table IVA

$a^2 I(h_m, \beta^2 g l^{-1} \nabla^2 h_m + l)$ in ft/hr for 1500 G.M.T. 27.1.1952

-12	-8	6	-17	-11	10	7	-3	-4	11	13	-2
3	7	7	-2	-13	-1	12	-1	-4	-20	-8	12
9	19	-1	-13	-11	-4	12	-15	12	-5	-33	-11
2	27	22	-24	-32	5	0	10	13	14	7	-15
6	-19	-18	-13	-4	-18	10	7	11	-11	5	-7
-15	0	-6	-24	-8	5	10	3	16	-16	-19	-23
2	-19	-30	-10	18	11	2	2	7	29	-5	28
-5	-8	-6	0	24	20	3	-1	6	21	-25	-22

Table IVB

$\frac{1}{3} a^2 I(h', \beta^2 g l^{-1} \nabla^2 h')$ in ft/hr for 1500 G.M.T. 27.1.1952

-1	-2	-2	0	2	2	1	-3	3	6	-2	-1
-1	-1	-2	0	1	2	2	-2	-2	1	-3	1
-2	1	0	-2	-1	0	0	-2	-1	-1	-2	-3
-1	0	0	-5	-3	1	0	-1	2	-2	4	-5
3	-3	-7	-5	1	3	2	2	3	-8	-3	-1
-4	-3	-3	-3	5	2	4	8	5	-4	-11	-2
-1	-2	-3	2	6	1	0	5	4	0	-4	-5
2	1	1	4	2	4	2	1	5	2	-4	3

Table IVC

$-\frac{8 a^2 \varpi_m \nabla^2 h'}{3(P_0 - P_1)}$ in ft/hr for 1500 G.M.T. 27.1.1952

-2	-2	0	-1	0	0	-1	0	-2	0	0	0
-3	-1	0	0	-1	0	-1	-1	-1	-2	-1	1
0	0	0	0	-1	0	0	0	-1	1	7	-2
0	-1	-1	1	-5	0	0	-1	-2	1	0	0
-1	0	0	-3	-3	0	1	0	-3	0	0	0
0	0	-2	-6	-4	0	-1	0	-8	0	3	-1
0	-2	-4	-5	0	1	-1	-1	-8	-10	1	0
0	-1	0	-1	0	0	0	-2	-2	-13	0	-2

/Table VA

Table VA

$a^2 J\left(h_m, \beta^2 g \ell^{-1} \nabla^2 h_m + \ell\right)$ in ft/hr for 1500 G.M.T. 20.12.1951

-59	23	68	48	25	13	37	-19	-32	9	15	3
-130	-37	53	56	0	-15	11	51	-21	-5	-3	-4
-18	-16	45	21	-23	16	-19	-1	-6	0	2	1
-26	-2	22	-5	-23	19	-8	0	-27	-8	-5	1
-7	-15	-10	-8	30	33	-3	6	-18	-1	-6	2
-5	-1	33	29	55	9	-5	-25	8	11	14	0
-8	1	-28	-13	-2	-7	-8	-12	2	18	14	2
-3	0	-18	35	14	-1	-2	0	-7	2	-7	-1

Table VB

$\frac{1}{3} a^2 J\left(h', \beta^2 g \ell^{-1} \nabla^2 h'\right)$ in ft/hr for 1500 G.M.T. 20.12.1951

-9	7	13	22	7	0	-4	-9	-2	3	2	-1
-7	5	8	10	-2	1	1	9	-10	2	0	-1
-5	3	7	-3	-3	1	-1	4	-4	-1	0	0
-2	3	1	-11	-6	1	-5	2	-3	-3	-1	1
2	4	0	6	3	3	2	2	-3	-3	0	1
3	0	3	7	11	-2	6	-8	-2	3	0	1
0	-2	-2	-11	3	-3	-5	-4	-2	4	0	0
1	0	-1	4	4	0	-4	0	1	-2	1	0

Table VC

$\frac{-8 a^2 \varpi_m \nabla^2 h'}{3(p_0 - p_1)}$ in ft/hr for 1500 G.M.T. 20.12.1951

26	-30	-10	-2	-3	-1	-6	-6	-1	-2	-1	0
8	-3	2	-5	-1	0	-8	-10	-1	-1	-1	-1
11	0	1	0	1	-1	1	1	-2	-2	-2	-1
2	-1	-1	0	4	-1	0	0	-4	-1	-1	-1
-1	-1	1	5	11	-2	-1	-1	-2	0	0	0
0	0	1	-7	-12	-5	1	-4	0	00	0	0
0	-1	0	0	-6	-4	0	1	0	00	0	0
0	-1	-2	4	-5	-2	1	1	1	-1	0	0

M. R. P. 790

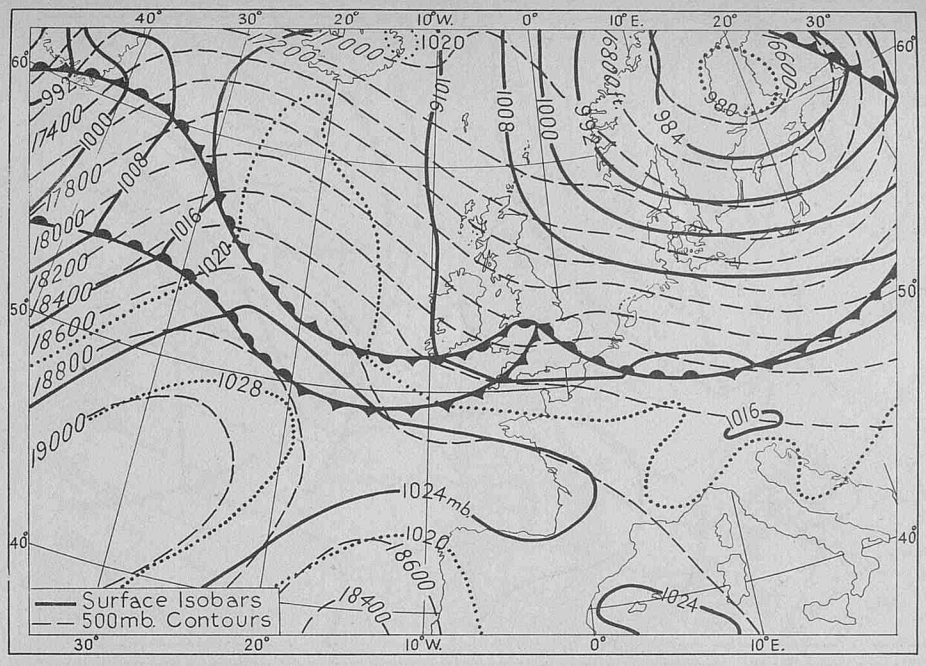

Fig. 1a. SYNOPTIC CHART 1500 G.M.T. 14. 3. 49

M. R. P. 790

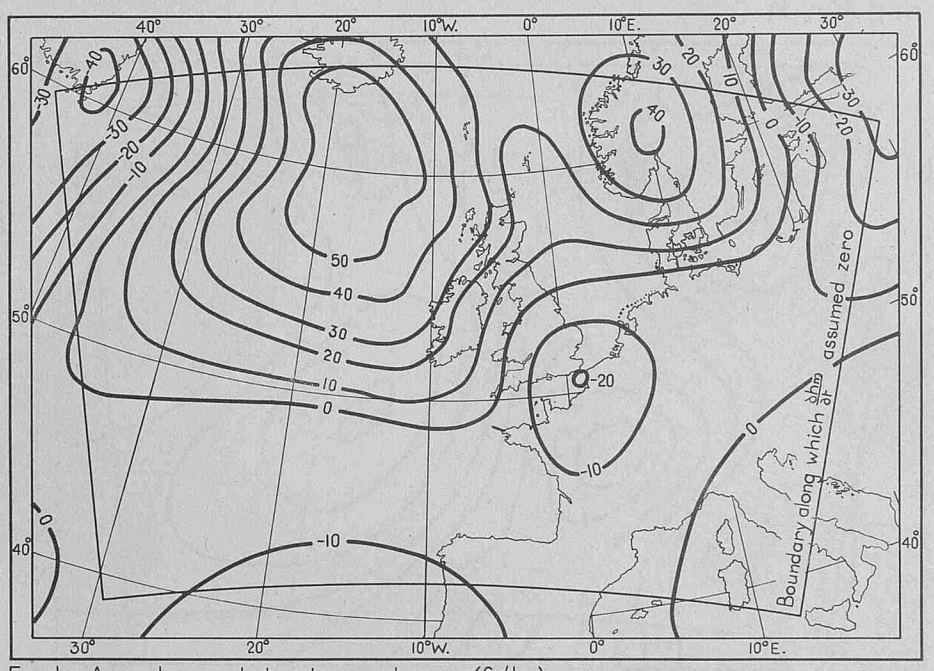

Fig. 1b. Actual 500-mb. height tendency (ft./hr.), 1500 G.M.T., 14.3.49

M. R. P. 790

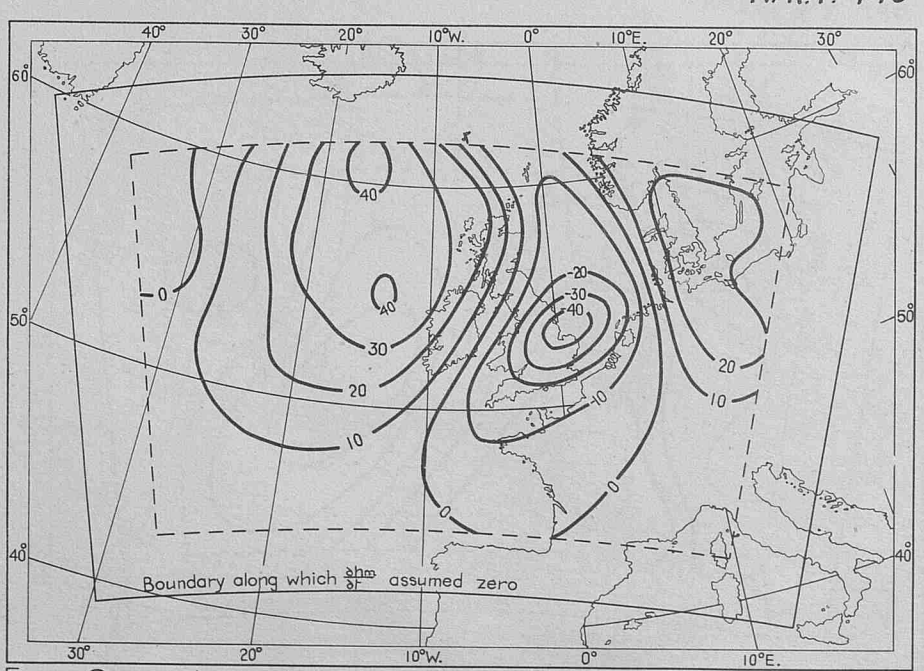

Fig. 1c. Computed 500-mb. height tendency (ft./hr.), 1500 G.M.T., 14.3.49

M. R. P. 790

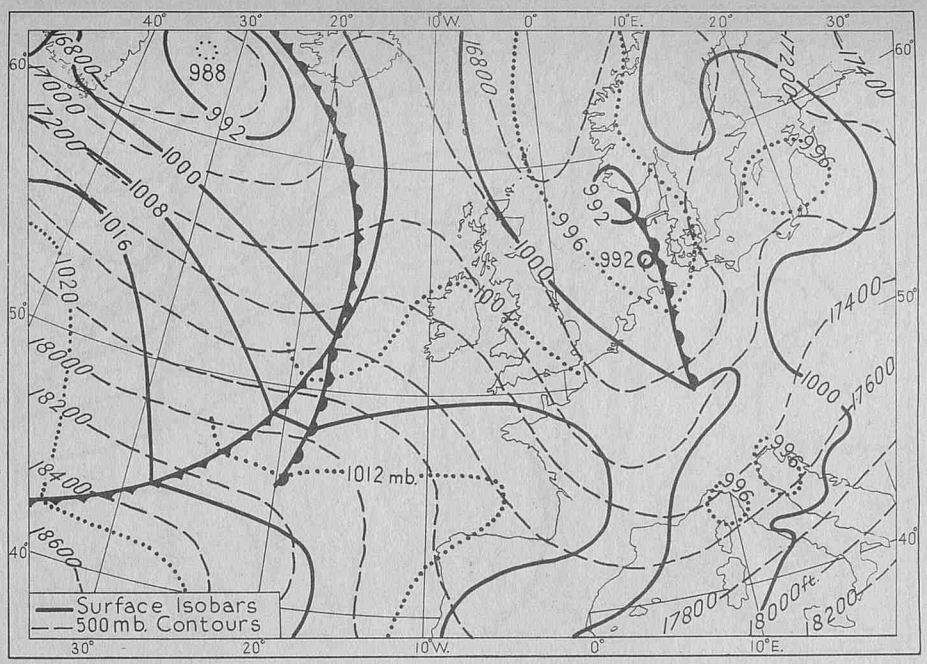

Fig. 2a. SYNOPTIC CHART 1500 G.M.T. 27. 1. 52

M. R. P. 790

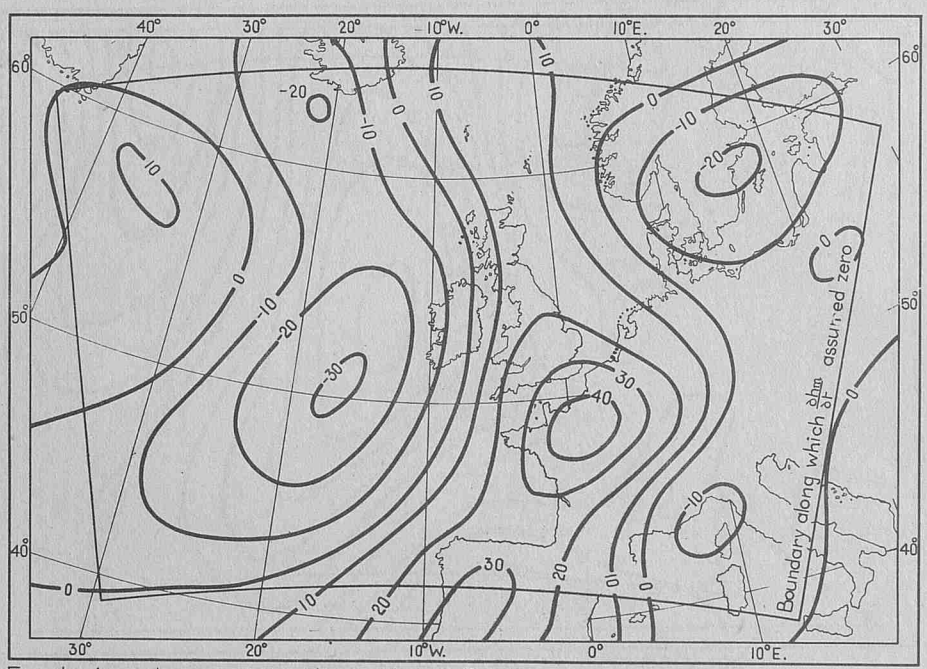

Fig. 2b. Actual 500-mb. height tendency (ft/hr.), 1500 G.M.T., 27. 1. 52

M. R. P. 790

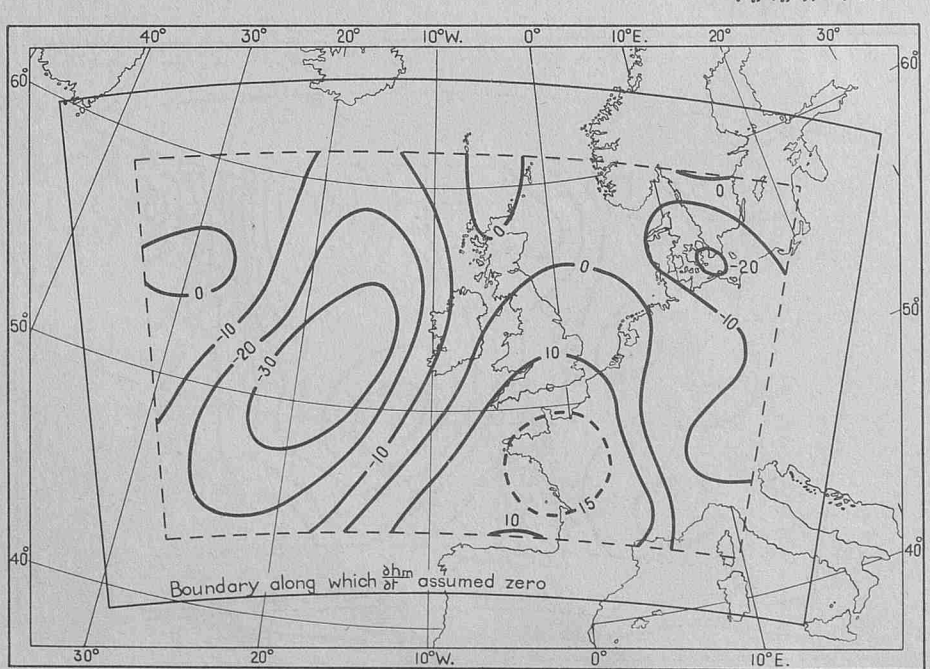

Fig. 2c. Computed 500-mb. height tendency (ft/hr.), 1500 G.M.T., 27. 1. 52

M.R.P. 790

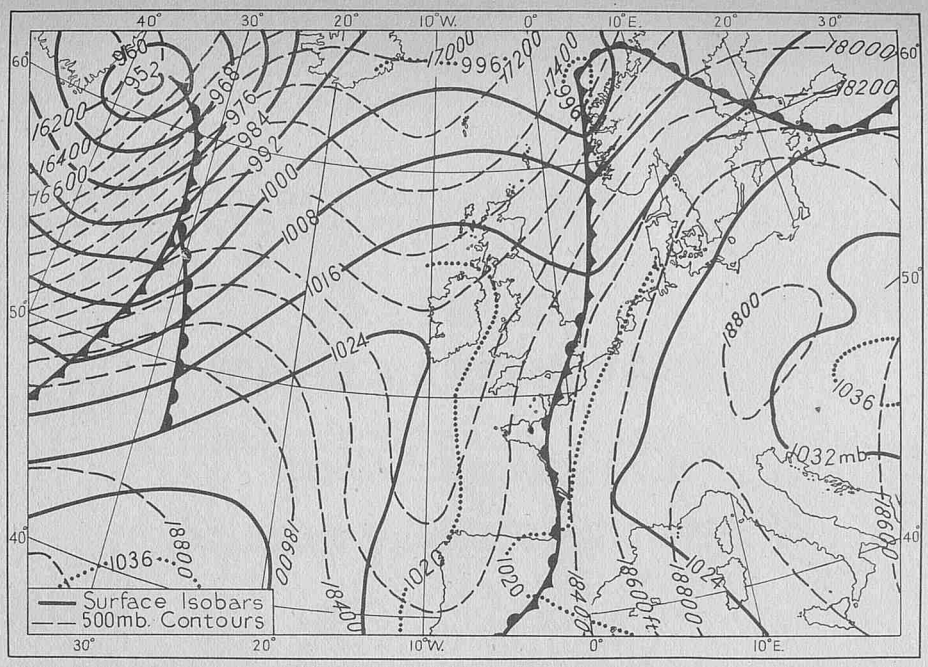

Fig. 3a. SYNOPTIC CHART 1500 G.M.T. 20. 12. 51

M.R.P. 790

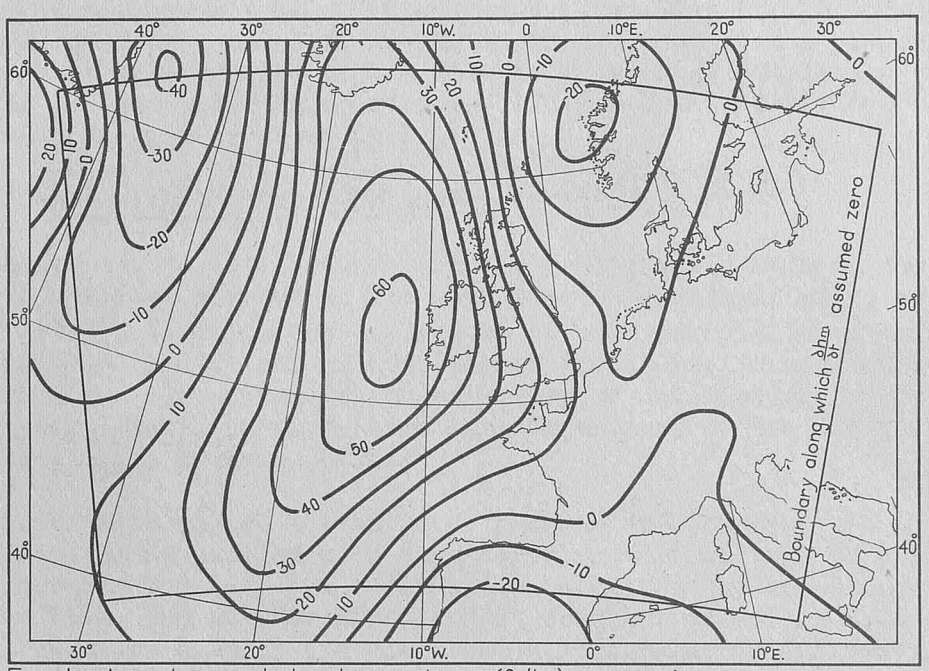

Fig. 3b. Actual 500-mb. height tendency (ft./hr.), 1500 G.M.T., 20.12.51

M.R.P. 790

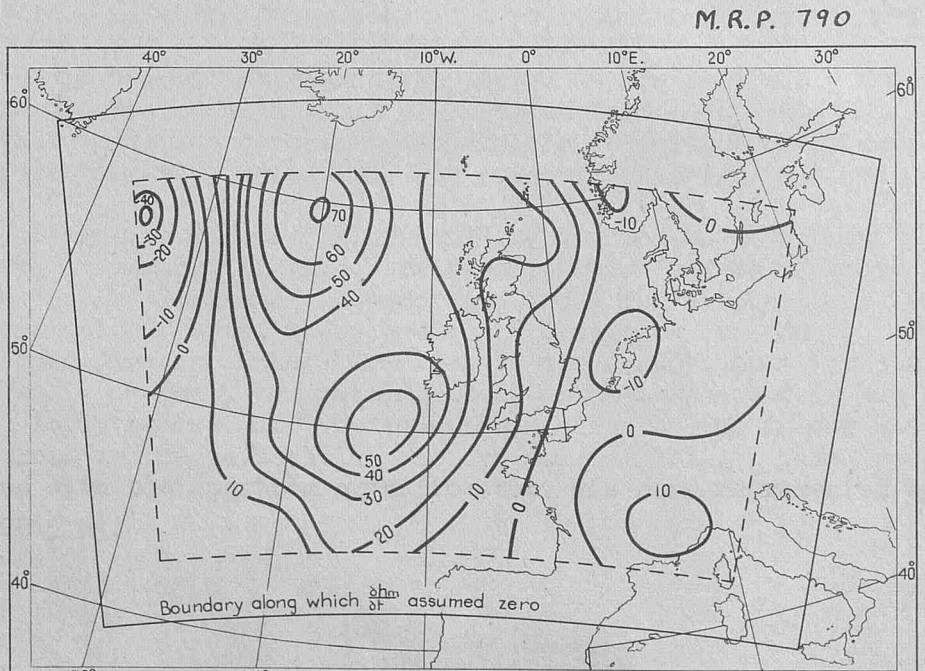

Fig. 3c. Computed 500-mb. height tendency (ft./hr.), 1500 G.M.T., 20.12.51

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 791 by Mr. M.J. Blackwell A copy is available in the Library of the Meteorological Office."

M.R.P. 791
S.C.I/75
24th March, 1953.

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On the development of an improved Robitzsch-type actinometer

by
M.J. Blackwell (Meteorological Office)

1. Introduction

The first mention of an actinometer of this type, by Robitzsch (1915), concerned an instrument working on the principle of two coaxial bimetallic rings, placed one above the other. By the year 1929, the mechanical design had been changed to an arrangement of two horizontal bimetallic strips, placed side by side. The instrument which is widely known and used at the present time is the Fuess-Robitzsch actinometer, with a detecting assembly consisting of a central black strip and two outer white strips. This model is described by Robitzsch (1932), and a diagram is given in Fig. 1.

2. Experience with the Fuess-Robitzsch actinometer

Extensive use of this actinometer was subsequently made on the continent, and several critical surveys of its performance published, e.g., by Morikofer and Thams (1937), in which the dependence of the sensitivity on position of the sun, intensity of radiation, ambient temperature, etc., was discussed. Consideration was also given to the variation of calibration factor with time of day or year, stability of factor, lag, importance of the paint used on the strips, etc., e.g., by Thams (1944).

The instrument has also been used in Japan and Canada. In the latter country, development has led to a considerably modified type in which a heavy bimetal increases the lag, but facilitates tabulation by smoothing the traces. At the same time, the stiffer bimetal has a smaller departure from the horizontal when irradiated, (a point which is considered further in this paper), and recording is carried out by using a vertical pen-arm and a drum revolving about a horizontal axis. (Ref. 5).

3. Experience with the Fuess-Robitzsch actinometer at Kew Observatory

A standard German instrument was tested at Kew in 1947. The records obtained were compared with those of a standard Moll Solarimeter, by estimating hourly totals and summing these to obtain the integrated daily totals of "Global" (sun plus sky) radiation, for each instrument.

The calibration factor, so obtained, was found to be within 1% of the figure given in the calibration certificate, which had been determined in Germany in 1942. Robitzsch however applied a correction term: $(1.3 - 0.005\ h)$ to the calibration factor, where h = solar elevation at midday. This term, if used, would give an annual variation in the factor about three times that found at Kew. In fact, the ratio of Moll to Robitzsch totals was constant for the period investigated, and the annual variation of the factor was deduced from the trend in the calibration factors for the Moll. This trend, based on comparisons with the Ångström pyrheliometer, was best represented by the term: $(1.1 - 0.0015\ h)$.

/There

There appeared to be some 10% fall in sensitivity between days of low, and days of high, radiation totals. The response at sunrise and sunset was slow. Links and pivots required attention, careful levelling was essential and pen friction was troublesome. The instrument was nevertheless regarded as suitable for recording radiation totals for periods of a day or longer; could be described as reasonably robust and responsive to semi-skilled attention, and had an all round accuracy of the order of ± 10%.

One further point of practical importance was that spot checks of the sensitivity, on the sun and against a standard instrument, were not possible. This was due to the slow response (15 to 20 min. for appreciable variations in intensity) and to the difficulty in shading the comparatively large sensitive surface, while keeping the portion of sky obscured to a minimum.

There remain two fundamental methods of calibration: comparison with a standard instrument over a long period, or a detailed laboratory calibration, using an artificial source. Both have been tried at Kew and found to be consistent, as will be shown. Once this consistency has been established, the laboratory method is to be recommended and a suitable routine technique is outlined below.

4. The Mk.I Meteorological Office bimetallic recorders

Six recorders were manufactured for the Meteorological Office by E.R.I.C. Ltd., of Acton, London. These were similar in design to the Fuess-Robitzsch. For the determination of their factors, a laboratory calibration was attempted, using a 2000 watt tungsten filament lamp as artificial source.

Preliminary ageing of the bimetals was carried out, by alternate heating and cooling in suitable temperature enclosures, before the arrival of the instruments at Kew in Feb. 1950. Subsequent ageing was carried out by exposing the painted surfaces intermittently to the radiation from the artificial source. The model under test was set up on an optical bench, with the black strip in the east-west position and the lamp due south. Some 20 min. were allowed for each reading, a further 20 min. being required to allow the pen to return to zero. Weekly clocks and drums were tried at first, but were later replaced by daily ones to reduce the sliding friction. The bearings, which were jewelled, were found to be easily damaged in transit, and were replaced by steel.

These precautions combined to reduce friction troubles and zero drift. Points were obtained for elevations of source from 15° to 90°, and calibration curves drawn, showing the sensitivity in mm. per cal./cm^2./min. The radiation intensities were measured by a calibrated Moll thermopile. A typical curve is shown in Fig. 2, the abscissa being in terms of sin. h, where h = elevation of the source. The factors were selected by averaging the sensitivities between 25° and 60°.

5. Development of the Mk. II recorder

The calibrations described above showed a marked variation of sensitivity with elevation, a 30% change occurring between 15° and 35°. It was realised that this was due, in part, to back scattering from the white pan lying beneath the bimetallic strips. This is brought out in Figs. 3(a) and 3(b). To reduce the effect, the spaces between the inner edge of the hemispherical dome and the outer edges of the white strips were covered in with suitably shaped pieces of white card. The idea had recently found favour elsewhere, see Götz and Casparis (1947).

The improved calibration characteristic confirmed the efficacy of such modifications, and a further improvement followed the extension of the white card to cover the white strips themselves. As a result, the variations of sensitivity, with both elevation and intensity of the incident radiation, were substantially reduced. Finally, it was found advantageous to remove

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the underlying white pan completely, to place the white strips on a lower plane than the black strip, as shown in Figs. 4(a) and 4(b), and to construct the side-shields of thin, white-painted aluminium.

Several minor modifications were introduced at this time, early in 1951. The former invar-brass bimetallic strip was replaced by "Morflex", a Wilco-Wiggin product which is thin, stiff, and has high transverse conductivity and high specific deflection. The time constant and sensitivity were both improved as a result of this change. Theoretical considerations also led to the rearrangement of the components so as to allow the pen-arm to be perpendicular to the spindle and, at the same time, perpendicular to the radius from the centre of the chart drum to the point of contact. This arrangement reduces pen-friction to a minimum.

6. Performance of the Mk. II recorder

The modified instrument was completed by May, 1951, and continuous records were obtained, at Kew, from then until Nov. 1951. The factor was ascertained by the laboratory method indicated above, and found to be unchanged within the experimental accuracy that could be attained at the time (2-3%), thus giving satisfactory evidence as to the stability of the instrument.

At the same time, fresh insight was gained from the results of a more detailed calibration, during which the instrument was placed on a turntable and calibrated for all angles of azimuth of source, as well as elevation. Sensitivity was least with the source due south and increased towards both east and west positions. This is due to the penetration of radiation on to the underlying white strip when the source is on the meridian (south). A more serious drawback lay in the marked asymmetry between the east and west sensitivities, resulting from the tilting of the black strip away from the horizontal as the ambient temperature varied. The effect is more noticeable at low elevations, and typical values may be quoted for 15°.

Azimuth	270	240	210	180	150	120	090
Sensitivity	61	45	29	25	17	21	28

The instrument was also tested for change of zero with ambient temperature, by placing it in an oven until a steady reading was obtained. A change of temperature of 10°F caused a movement of about 1 mm. in the position of the zero. This may be compared with the effects of friction, which may cause zero errors of the order of 2 mm.

The six months of daily records were analysed by measuring the areas under the traces with a planimeter, the zero line being taken as the lowest level on the chart reached by the pen during the night. These areas were correlated with the daily totals of radiation given by a Moll thermopile which registered automatically via an integrating motor and counter. The histogram of the six months' observations, see Fig. 5, shows a fairly sharp peak with a mean deviation of ± 10%. From this, the factor of 16.2 cal. per sq. cm. of chart area was deduced.

A further point elucidated by the analysis concerns the variation of sensitivity with temperature. This is generally estimated to be about 0.5% decrease per 1°F, see e.g. Ångström (1951). In the current investigation, a parameter, $\frac{1}{2}$ $(T_{0900} + T_{1500})$, where T is the screen temperature, was taken as representative of daytime conditions. Monthly means of this parameter were plotted against the mean factors for each month, see Fig. 6. From the slope of the graph, a 0.25% increase per 1°F was obtained, but it must be remembered that the variation of sensitivity with elevation and azimuth is included in this coefficient.

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To compare the two methods of calibration, sensitivities were computed from the laboratory results by plotting interpolated values against time of day (L.A.T.), at two-hourly intervals, for four representative days: the summer and winter solstices and the equinoxes. The factors obtained in this way were: Spring, Summer, Autumn - 36 mm. per cal./cm^2./min., Winter - 25 mm. per cal./cm^2./min. The corresponding mean factor for the six months June to November is found to be 33 mm. per cal./cm^2./min. which for one hour, (11.4 mm. on the chart), gives:

$$3.76 \text{ cm}^2. \equiv 60 \text{ cal.}$$

$$\text{or} \quad 1.00 \text{ cm}^2. \equiv 15.9 \text{ cal.}$$

This result must be compared with the mean value, given by direct comparison, namely:

$$1.00 \text{ cm}^2. \equiv 16.2 \text{ cal.}$$

7. Development of the Mk. III recorder

The latest version of the bimetallic actinometer took shape during the first half of 1952, and was designed to eliminate the most serious of the short-comings indicated by the previous work. The foremost of these, the asymmetry of the east and west sensitivities, was overcome in the following manner.

Two further bimetallic strips were used to support the assembly of black and shaded strips, and were half the length of the sensitive strips. Let us suppose that, for a given rise in temperature, the upper ends of these supports are now inclined at an angle $\theta°$ to the vertical. Then the two ends of the sensitive strips will be inclined at the angle $\theta°$ to the horizontal, and the tangent to the midpoint will be itself horizontal, see Figs. 7(a) and 7(b). The system is self-compensating for all changes of temperature when the assembly has once been levelled at any arbitrary temperature. The compensation does not apply to the attitude of the black strip when this is strongly irradiated, but is effective when most requred, i.e.: at dusk and dawn, when the elevation of the source is lowest.

So that zero troubles should not be introduced, care was taken to arrange these supports to be in line with the link rod between the black strip and the crank on the spindle. A stiffer bimetal, "Highflex 45", was used in place of the "Morflex", and the strength of the system was improved by widening the black strip and reducing the width of the two shaded strips. This equalises the load between them, and can be safely done, provided the ratio of width to length is kept below $1/4$ to prevent appreciable transverse bending. Previously all three strips were of identical dimensions.

The residual variation of sensitivity with azimuth was largely removed by adding flanges to the side-shields, and reducing the clearance between dome, side-shields and black strip. This isolates the cool strips from radiation at all angles of elevation and azimuth. Two capstan screws, operating on two U - pieces, enable the assembly of strips to be levelled before use. Plate 1 shows the finished model, with the outer case removed.

8. Performance of the Mk. III recorder

Fig. 8 shows the characteristics of this model superimposed on those of the Mk. II recorder. The sensitivities have been reduced to the same scale for ease in comparison. Not only has the asymmetry been almost completely eliminated, but the amplitude of the variations with azimuth has also been considerably reduced, being little more than the limits of accuracy attainable by the laboratory method of calibration. The sensitivity at low elevations (say 15°) is now rather higher than before, due to the improved shielding afforded by the flanges.

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The linearity of response, with intensity of radiation, was investigated and found to extend to 1.3 cal./cm^2./min., an intensity rarely exceeded in this country. A final test was made to calculate the lag of the instrument by measuring the difference in temperature between the black and cool strips, when the instrument was suddenly exposed to incident radiation. The intensity used was 0.4 cal./cm^2./min. Fig. 9 shows the exponential approach to the steady state, the time constant being about 4 min. Owing to frictional processes, the time constant found in practice is about 7 min.

The instrument was used to obtain daily records throughout May, 1952. When compared with the Moll thermopile and integrator, as before, the scatter was found to be less than $\pm$ 5%, as shown in Fig. 10.

9. Routine calibrations

With detailed knowledge of the Mk. III instrument's characteristics, it becomes possible to eliminate much of the work outlined above, and to develop a routine method which can be completed in a fraction of the time. This is based on the validity of two assumptions: (a) The factor, computed from the results of the detailed laboratory calibration, agrees with the figure, obtained by comparison with another type of instrument, over a period of six months. (b) Because of the negligible variation of sensitivity with azimuth, the sensitivity found for the south position may be taken as representative of all positions for that particular angle of elevation.

The instrument is therefore placed on an optical bench, with the artificial source (a 2000 watt, 220 volt, tungsten-filament lamp) in the south position. The lamp is under-run at about 180 volts to improve the stability of its output. Measurements are taken at 15° intervals of elevation, from 15° to 90°, using a Moll thermopile as reference. The lamp is switched off between readings, to allow the pen to return to zero each time. From these points, a graph of sensitivity against elevation is constructed.

When the latitude of the site at which recording is to take place has been ascertained, the elevation of the sun is computed, for two-hourly intervals, for the equinoxes and the summer and winter solstices. The corresponding sensitivities are obtained for these representative days by interpolation, and hence the factors are deduced for each season of the year.

10. Conclusion and suggestions for future work

In the field, this actinometer will give a reasonable indication of that part of the fine structure of the intensity variations which is of period greater than 5 min., and will give daily totals of Global (sun plus sky) radiation with an accuracy of the same order as that of the Moll thermopile. It is desirable that comparisons with the standard Moll solarimeter be made over a period of twelve months to assess the stability of the calibration factor under working conditions. Finally, it is suggested that any future work should include an investigation of the properties and durability of suitable paints for the black strip and the side-shields.

11. Acknowledgments

I wish to express my thanks to Dr. G.D. Robinson for much helpful discussion of the work, and to Mr. R.H. Eldridge (now at London Airport) who did the earlier laboratory work and was responsible for the development of the Mk. II recorder. Thanks are also due to Mr. A.J. Lander for advice in the construction stages, and to Mr. D.B.B. Powell for assistance in calibration, and analysis of the records.

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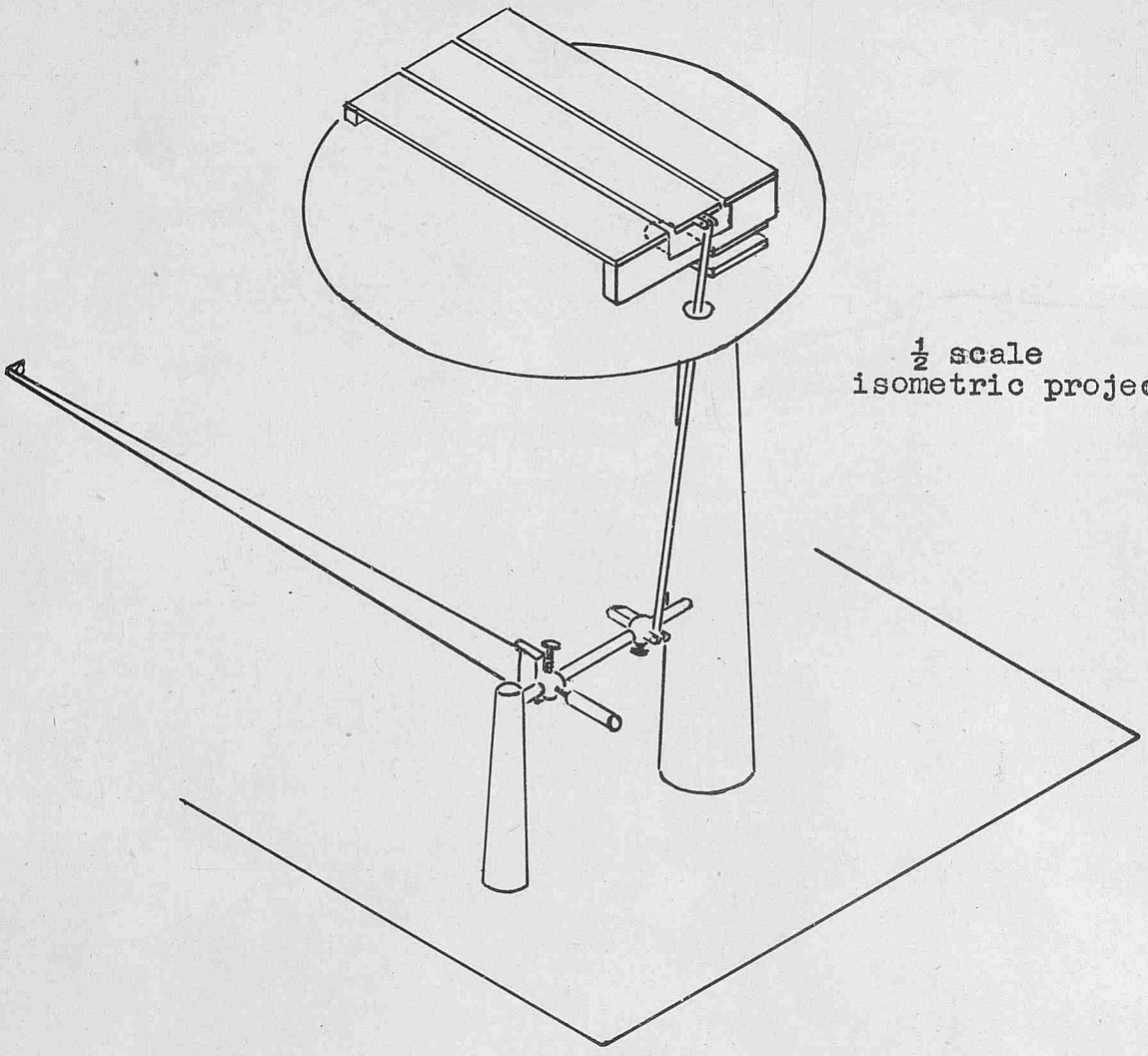

½ scale
isometric projection

Fig.1
Diagram of the Fuess-Robitzsch actinometer

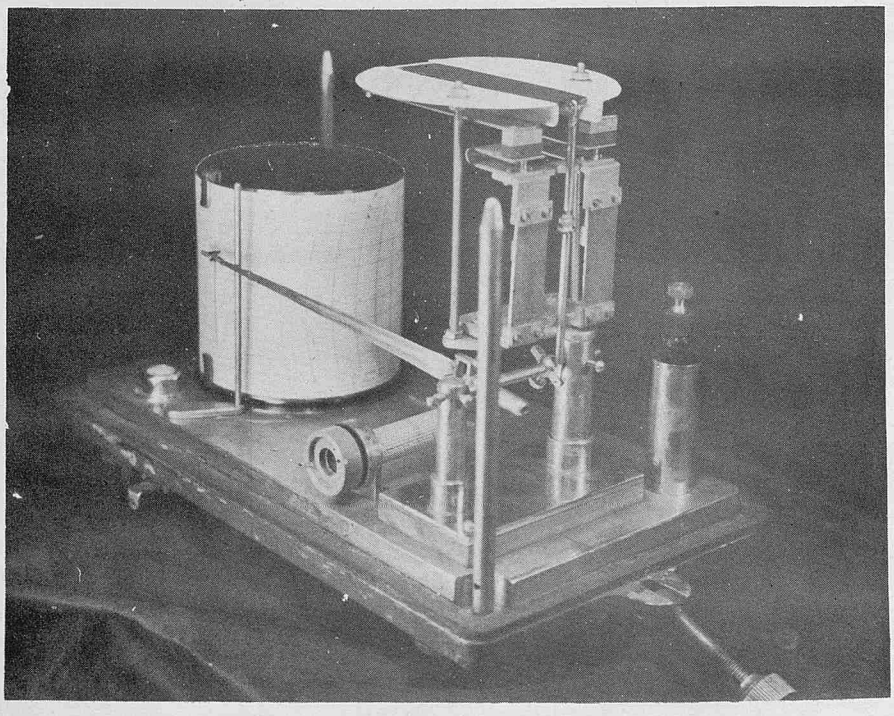

Plate 1
The Mk.III recorder

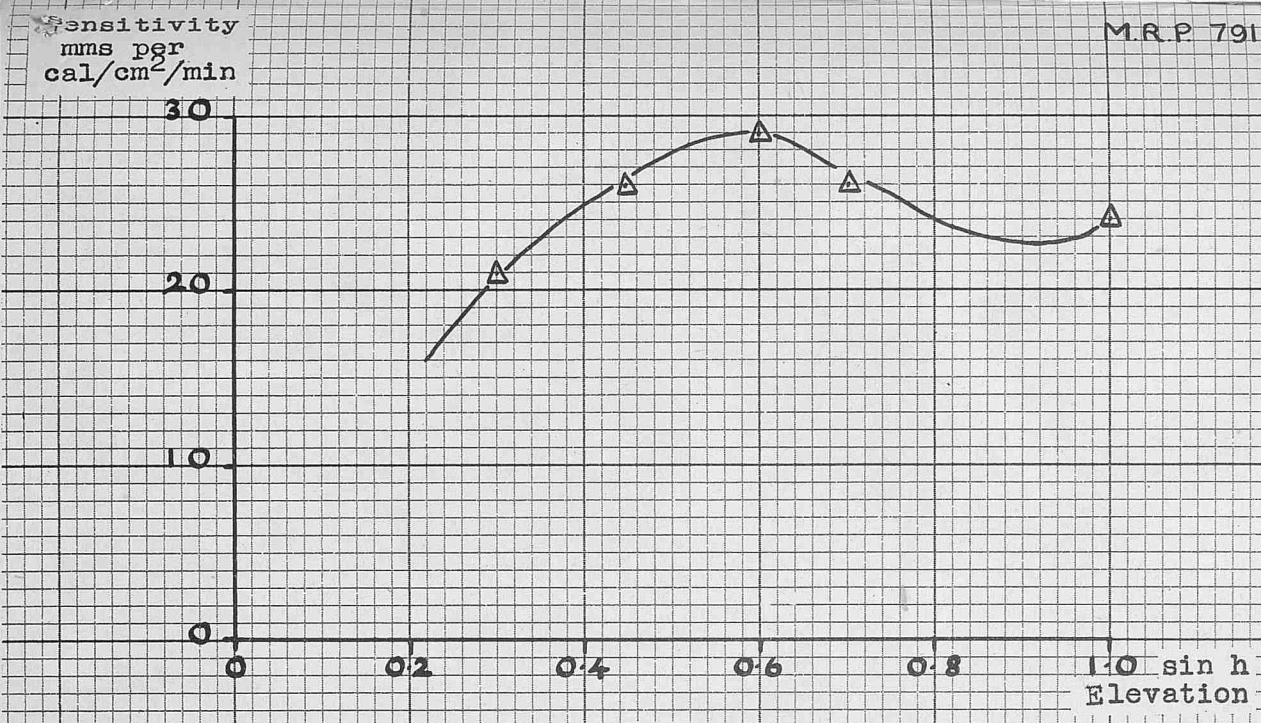

Fig.2
Typical calibration curve of
the Mk.I recorder

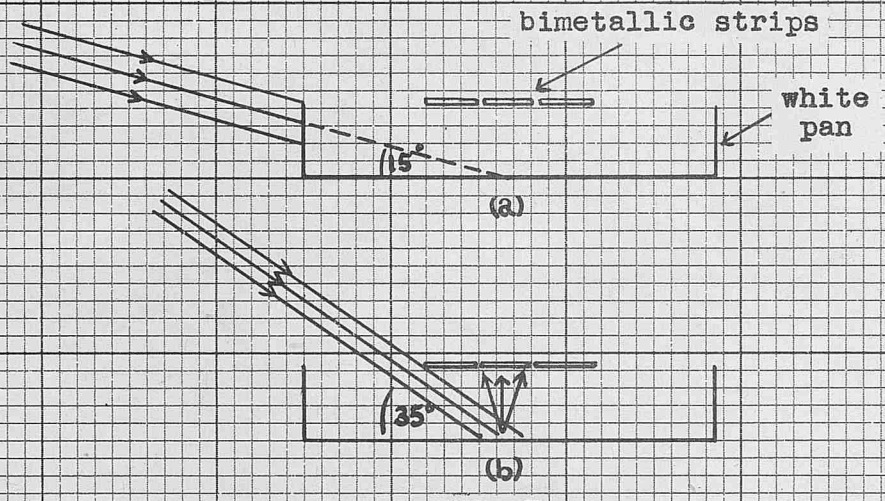

Fig.3
Variation of back scattering with
elevation of incident radiation

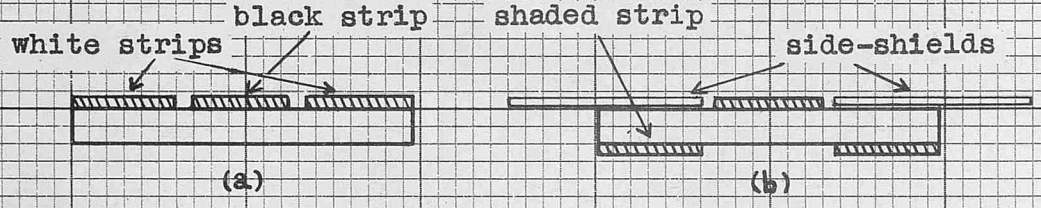

Fig.4
Introduction of side-shields

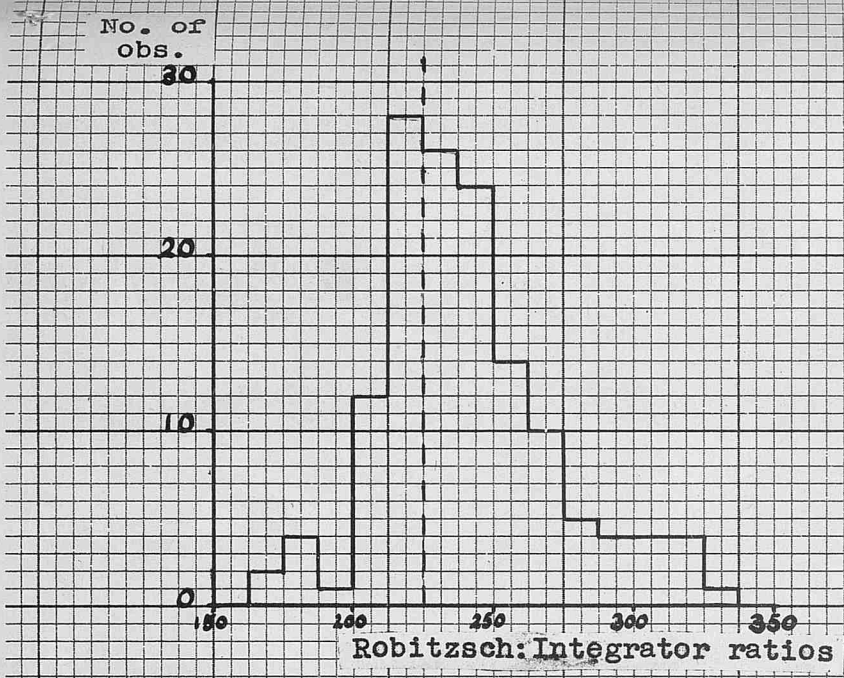

Fig.5
Histogram of six months of Robitzsch: Integrator ratios

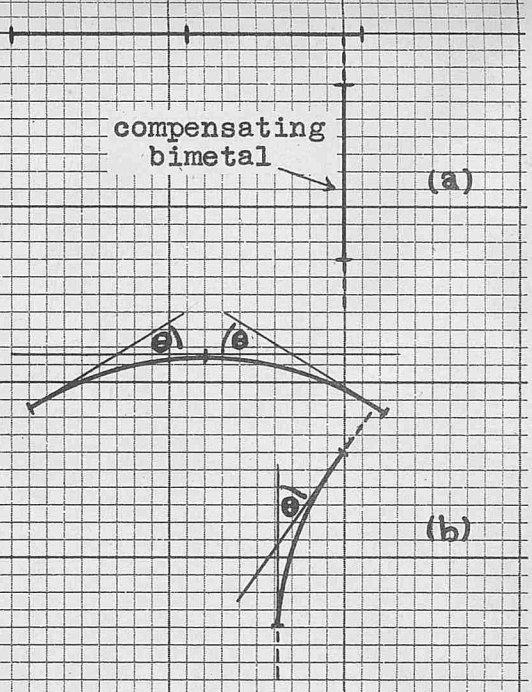

Fig.7
Principle of self-compensating tilt mechanism

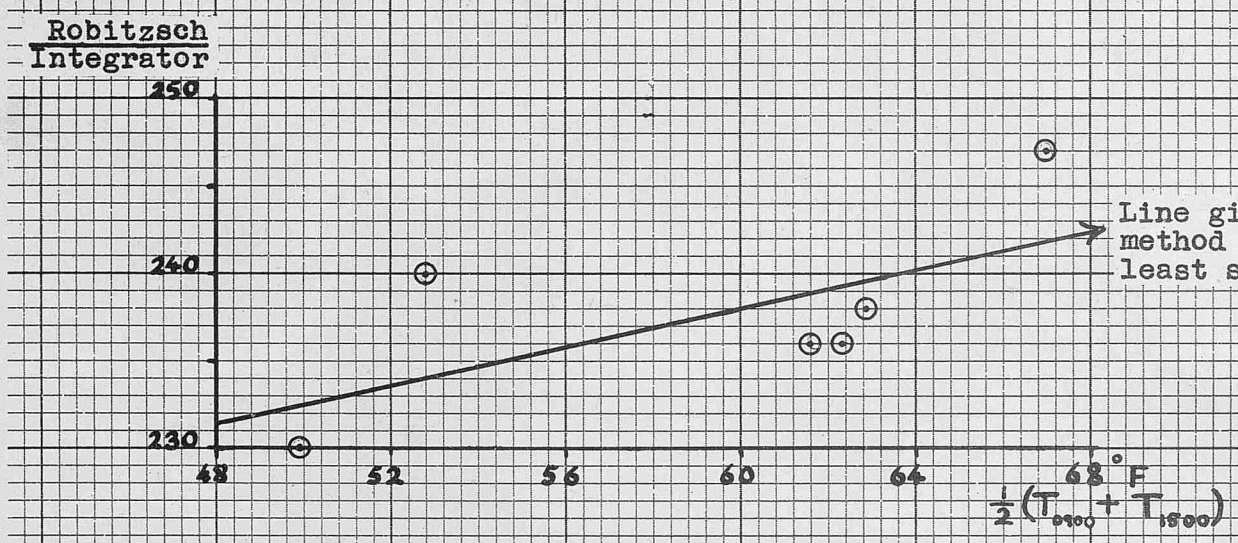

Fig.6
Sensitivity as a function of temperature (including effects of elevation and azimuth of the sun)

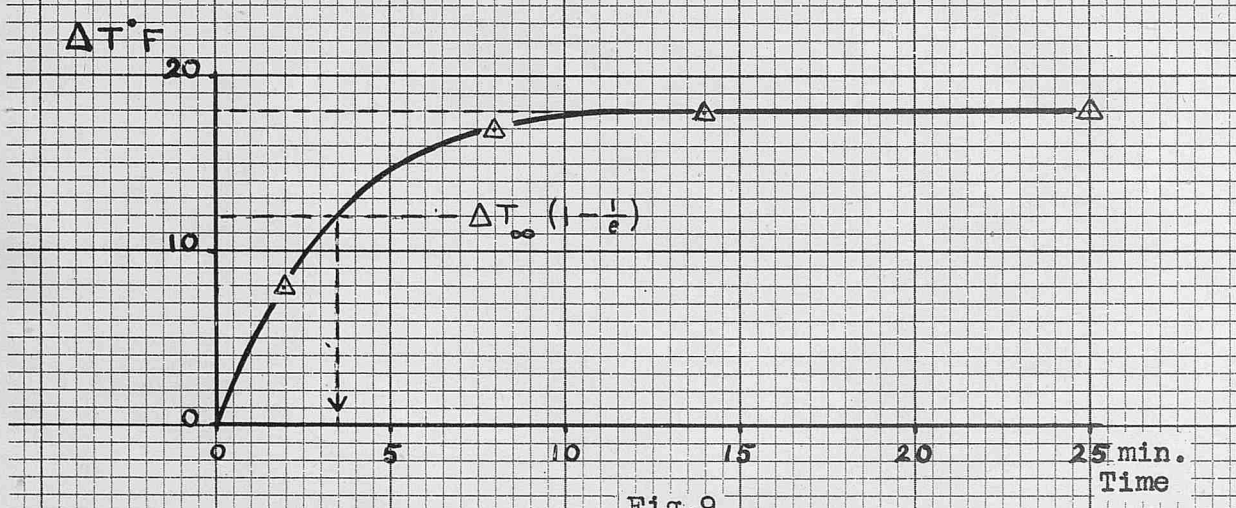

Fig.9
Time constant of the Mk.III recorder

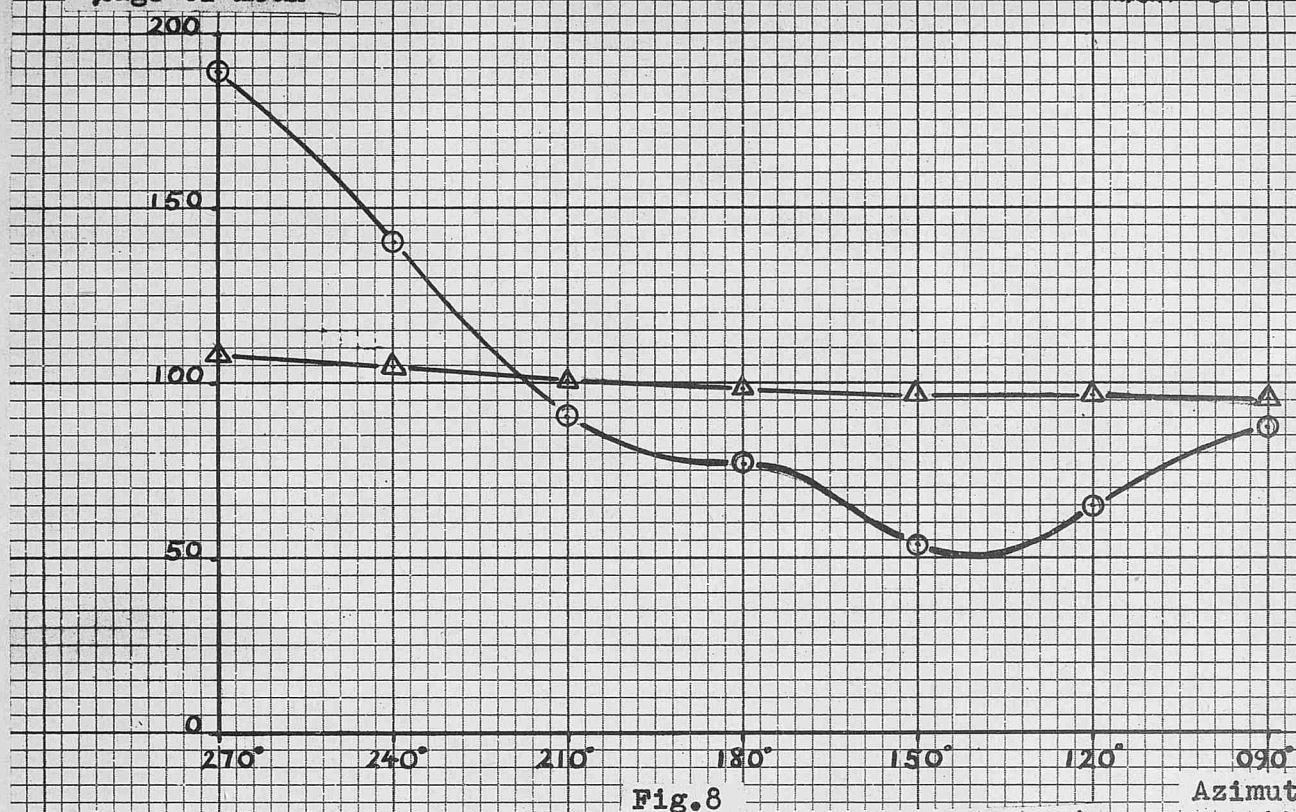

Fig.8
Comparison of characteristics of the Mk.II (⊙---⊙) and Mk.III (△---△) recorders for elevation 15°

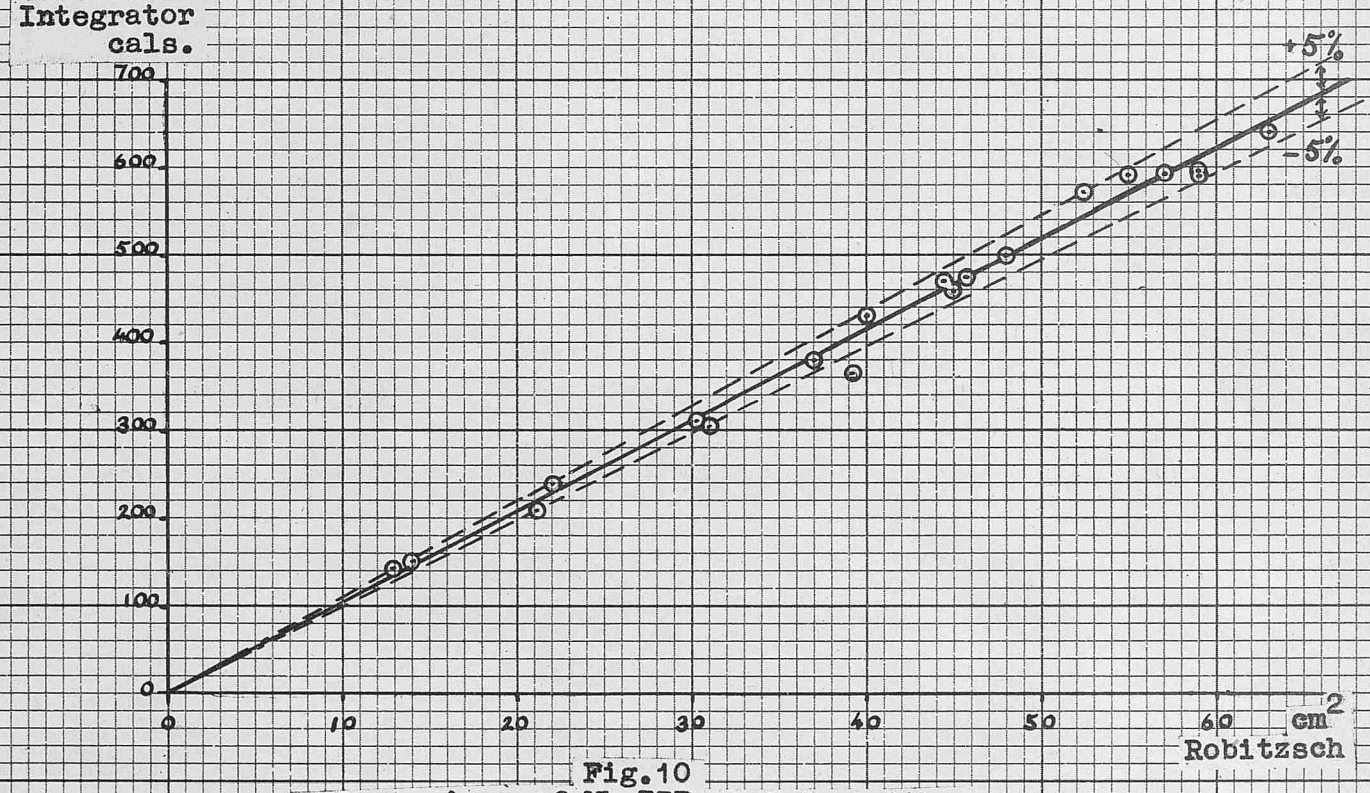

Fig.10
Comparison of Mk.III recorder with Integrator for all occasions, May 1952

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No ..... by Mr.....
A copy is available in the Library of the Meteorological Office."

M.R.P. 792

S.C.III/148

26th March, 1953

AD 32644

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An Account of One of the Methods Employed in the Investigation of Clear Air Turbulence

by

D.G. James (Meteorological Office)

1. Summary. A description is given of one of the methods now used at the Meteorological Research Flight to investigate turbulence in the free atmosphere. Measurements of wind together with horizontal and vertical wind shear are made by discharging lines of smoke puffs over a camera obscura at East Hill. Rapid fluctuations of temperature and gustiness are obtained by the use of an ultra rapid thermometer, a hot wire anemometer, and an accelerometer and are photographically recorded. Readings of the height and airspeed together with the mean values of temperature and humidity are also made.

A detailed account of the first experiment of this type made on the 22nd May 1952 is given. The results show a good correlation between the degree of turbulence and the Richardson number. Attempts to correlate the hot wire anemometer records with those of the ultra rapid thermometer are also described.

2. Introduction. Several flights have been made by a Hastings aircraft of the Meteorological Research Flight to find and examine the properties of clear air turbulence. Unfortunately the frequency of occurrence of this phenomenon below 25,000 feet, the ceiling of the aircraft, is extremely low, if it is confined to occasions when noticeable bumpiness is experienced by the passengers in the aircraft. It is thought, however, that this bumpiness is caused by turbulence of a scale closely related to the aircraft size. Turbulence of a smaller scale, and indeed of a much larger scale, is considerably more frequent. (Ref.1).

Following continual failures to find turbulence of the critical size, it was decided to investigate smaller scale turbulence. Consequently, a random block of cloudless air some 2,000 feet thick, in the free atmosphere, was examined to determine mean temperatures, small scale temperature fluctuations, frost points, mean winds and small scale wind fluctuations.

These observations determine to a large extent the character of a limited air mass, and most of the parameters relative to its behaviour. The wind shears in the vertical and horizontal planes are readily found, and the Richardson number between two levels may also be calculated. Any agreement between the scale of the small fluctuations of temperature and wind and the value of the Richardson number is at once evident, even though the vertical shear is obtained by calculating the mean wind at each level.

Similar work has been carried out by Bannon, (1951), who evaluated the vertical and horizontal shears from corresponding radar winds at different radio-sonde stations in Great Britain. He found some correlation of the Richardson number with reports of bumpiness from Meteorological Research Flight and other aircraft.

3. Flight details, taking a typical sortie as illustration.

The crew of the Hastings aircraft consists of a pilot, co-pilot, engineer, and five observers. These observers record readings of the flat plate thermometer, frost point hygrometer, height, airspeed, and operate the photographically recording ultra rapid thermometer and hot wire anemometer, whilst firing smoke puffs when required.

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On the 22nd May 1952 an anticyclone to the west of Ireland was intensifying and moving slowly north-east, giving a north-easterly airstream over southern England. A small layer of broken altocumulus at 12,000 feet made it necessary for the top run to be made at 11,000 feet, from which height the aircraft was always clearly visible on the camera obscura table. Pairs of runs were then carried out at 500 ft. intervals down to 9,000 feet, the aircraft flying crosswind, during which time a total of 57 puffs was fired. Generally these were fired in bursts of six at ten second intervals, i.e., at a distance of some 2,000 feet between two consecutive puffs. Two such bursts were fired at each level, though in some cases odd puffs were missed because of faulty cartridges. The path of each puff was traced on the camera obscura table at East Hill, and its position recorded every ten seconds for two minutes. It was necessary to allow the ground observers sufficient time to plot these positions on the table, and consequently a time interval of two or three minutes was allowed to elapse between each run.

The runs at each level were duplicated so as to permit records of small scale fluctuations of temperature and wind to be obtained consecutively. The traces of the ultra rapid thermometer and the hot wire anemometer used for this purpose are recorded on the same camera, but a simultaneous record is impossible at present. This means that we cannot correlate any individual small scale temperature variation with a wind gust as measured on these traces. However, we can at least assume that no drastic change has occurred during the two or three minutes time lapse between each run, and that the traces are representative of the character of the air mass at that particular level.

The ultra rapid thermometer and hot wire anemometer runs were each of two minutes duration, the recording camera being switched on when the first puff was fired. This was necessary since the aircraft could not be seen for the whole of its run, so that the ground controller gave the signal for the first puff to be fired as the aircraft appeared on the table. Following the run, the camera was switched off while the aircraft turned on to the reciprocal course along the same straight line as nearly as could be determined by the air position indicator.

4. Instruments.

(a) Aircraft instruments.

(i) The flat plate thermometer, which, when corrected for height and airspeed, gives the mean temperature at each level, and also any gradual variation from the mean. This instrument has a lag of about eight seconds.

(ii) The Dobson-Brewer frost point hygrometer, which is slow and has an accuracy of about $\pm$ 1°F, at a frost point of 10°F to 20°F.

(iii) The ultra rapid thermometer as described in M.R.P. 658 (Grant, 1951). Small rapid fluctuations from the temperature mean, whether positive or negative, are recorded by photographing on 35mm. film the movement of a spot on a cathode ray tube. It is possible to ascertain, by correct setting of the controls, which are the positive and which the negative fluctuations, and to interpret the films accordingly. The lag of the instrument is of the order of four milli-seconds, whilst variations of 0.1°F can be accurately read.

(iv) The hot wire anemometer. Small scale fluctuations of wind relative to the motion of the aircraft are at present recorded by the same electronic equipment and camera as is used for the ultra rapid thermometer. It is possible to observe which are the increases and which the decreases of the incident airstream, but with the design of hot wire anemometer used on the sortie we cannot distinguish between

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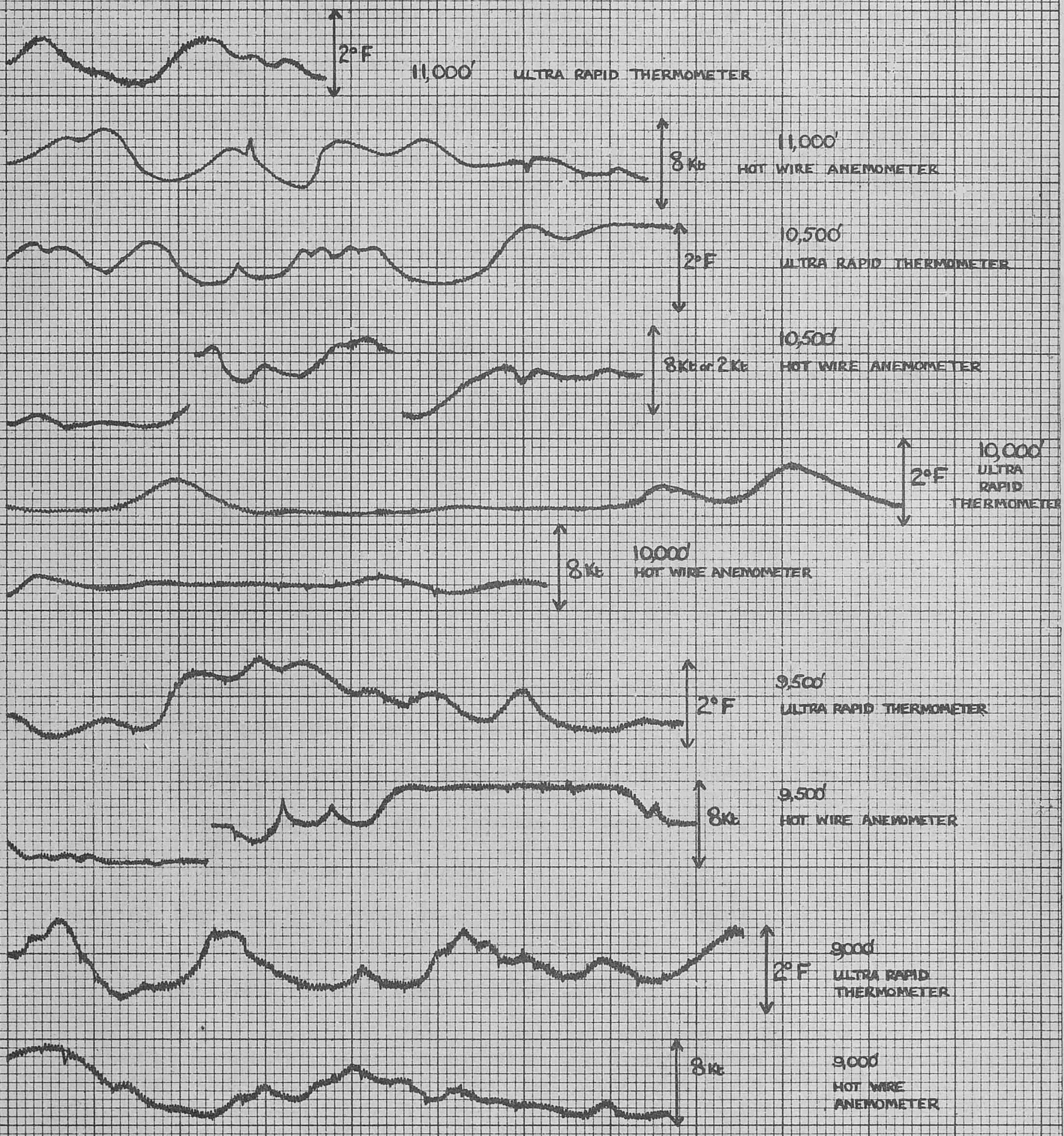
M.R.P. 792.

Instrument Traces From Turbulence Investigation on May 22nd, 1952



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upward and downward gusts, or in fact gusts coming from any direction in planes containing the line of flight of the aircraft. (It might be mentioned here that there is now at the Meteorological Research Flight an instrument which measures vertical gusts only, with sense, but which has not yet been used on this work.) However, the record will certainly give us an idea of the amount and type of velocity fluctuation at any level, and will enable a comparison to be made with the variations at other levels. The lag of the instrument is of the order of 20-30 milli-seconds, whilst variations of $\frac{1}{4}$kt. can easily be read.

(v) The accelerometer, which is photographically recorded and has a full scale reading of $\pm$ 0.25 g. Unfortunately, owing to a fault which developed during the flight, the film for this instrument is not available for this particular sortie.

(vi) Height and airspeed are read from standard aircraft instruments the former being set to the ICAN pressure of 1013 mb.

(b) Ground instrument.

The camera obscura table employed has an area of vision of four miles by four miles for an aircraft flying at 10,000 feet. The images of the puffs are plotted every ten seconds for two minutes, and the final results are given with an accuracy of 0.01 inches and 0.1° for each position of the image. The wind velocity can be obtained from the formula

$$V = \frac{h\,d}{302} \text{ ft./sec.,}$$

where 'h' is the height of the aircraft in feet, and 'd' is the ten-second displacement of the puff on the table in inches.

5. Information available from the Camera Obscura Measurements.

From the above instruments we can obtain:-

(a) A wind velocity - space profile in the horizontal at each level.

(b) A wind velocity - space profile in the vertical.

(c) A mean wind at each level.

(d) The mean horizontal wind shear along the X and Y axes, i.e., east-west and north-south, in the horizontal plane at each level.

(e) The mean vertical shear between levels.

Since it is not possible to fire the lines of puffs vertically above one another at each level, - such a procedure would require the use of a precision navigator, - we cannot obtain the vertical shear in a given vertical plane. Even if such equipment were available, the time interval between the puffs at consecutive levels, which is of the order of four minutes, would render the results objectionable from this point of view. An improvement would be effected by the use of two aircraft flying vertically above one another and firing puffs at the same instant. Such a scheme is being considered at present at the Meteorological Research Flight.

6. Analysis of the film records.

A complete analysis of each trace has been made. A short summary is given below for both the ultra rapid thermometer and the hot wire anemometer. Since the quality of the prints of these records is not good, traces of the films have been made and are presented at the end of this paper. It is important to realise that for the hot wire anemometer records, the fluctuations are

/present

present together with any change in airspeed of the aircraft. The fluctuations are obtained approximately by plotting on the film the spot readings of the airspeed made every ten seconds, and subtracting ordinates. Generally, however, this procedure is not necessary since the gusts encountered have edges sharper than the response of the airspeed indicator, and can easily be picked out from any slow change of aircraft speed. The pilot is instructed to maintain a constant indicated airspeed of 120 kt. for the hot wire anemometer, and constant height for the ultra rapid thermometer.

(1) 11,000 feet.

Ultra rapid thermometer.

Duration of trace:- 1 min. 40 sec., or 20,000 ft airspace. Sensitivity:- 2°F. full scale.

The first part of the film is fogged, but on the remainder we have two large full scale peaks, the trace falling from the first maximum by 1°F in 3,000 feet, and from the second by 0.5°F in 1,500 feet. There are numerous smaller fluctuations, all of which are not reproduced in the trace, of the order of 0.2°F which last over 30 feet or so, but appear quite random in relation to the larger fluctuations. The trace is noisy.

Hot wire anemometer.

Duration of trace:- 1 min. 50 sec., or 22,000 ft airspace. Sensitivity:- 8 kt. full scale.

Gusts of the order of four to six knots are evident on this trace. Generally the changes are slow, being of the order of three to four knots in 3,000 feet, but there is one rise of four knots in 1,000 feet. The latter half of the trace indicates a slow decrease in aircraft speed, with gusts superimposed. The trace is noisy.

(2) 10,500 feet.

Ultra rapid thermometer.

Duration of trace:- 1 min. 55 sec., or 23,000 ft airspace. Sensitivity:- 2°F full scale.

The total range of temperature variation encountered on this run is 1.3°F. Between these extremes there are numerous large and small scale fluctuations of 0.9°F to 0.3°F. Generally these are of long period i.e. greater than 10 sec. flying time, with shorter period (i.e. less than 5 sec. flying time) smaller fluctuations superimposed. The trace is noisy.

Hot wire anemometer.

Duration of trace:- 1 min. 45 sec., or 21,000 ft airspace. Sensitivity:- 8 kt. full scale, but at the centre of the film the sensitivity is increased to 2 kt. full scale.

The trace is relatively quiet for the first 30 sec., there being only small gusts of the order of 0.5 kt. The sensitivity is then increased and rapid fluctuations of the order of 0.2 to 0.6 kt. are evident with little change in aircraft speed. The original sensitivity is now restored, but there are still only minor fluctuations of the same order. The trace is quiet.

(3) 10,000 feet.

Ultra rapid thermometer.

Duration of trace:- 2 min. 40 sec., or 32,000 ft. airspace. Sensitivity:- 2°F full scale.

/This

This trace is particularly free from the smaller temperature fluctuations. There are, in fact, only three well defined temperature changes of long period. These are respectively, 0.6°F over 3,600 ft., 0.5°F over 2,200 ft., and 1.0°F over 4,000 ft. The trace is very quiet.

Hot wire anemometer.

Duration of trace:- 1 min. 45 sec., or 21,000 ft. airspace.
Sensitivity:- 8 kt. full scale.

After seven seconds the spot was moved to a more convenient position on the cathode ray tube, but only small gusts of the order of 0.5 kt. to 1.0 kt. are observed. The trace is very quiet.

(4) 9,500 feet.

Ultra rapid thermometer. Duration:- 2 min, or 24,000 ft. airspace. 2°F full scale.

The total range of temperature variation is 1.6°F. The largest change occurs after 25 sec. of the trace when there is a sudden rise of 1.3°F in 1,000 feet. The temperature falls gradually from this value until, at the end of the trace, some 17,000 feet later, the temperature has fallen to its original value. Superimposed on this general decrease are numerous smaller fluctuations of the order of 0.3°F to 0.5°F, and of varying periods. These are similar to the changes which occurred in the trace before the sudden rise. The trace is noisy.

Hot wire anemometer. Duration of trace:- 2 min, or 24,000 ft. airspace. Sensitivity:- 8 kt. full scale.

This trace can be divided into three parts. The first and third indicate that a change of airspeed has occurred between them, even though they show little small scale fluctuations. The intervening period, however, - some 40 seconds - is quite noisy, there being several sharp edged gusts of the order of two to four knots. The trace is noisy.

(5) 9,000 feet.

Ultra rapid thermometer.

Duration of trace:- 2 min, 5 sec., or 25,000 ft airspace.
Sensitivity:- 2°F full scale.

The largest variation occurs soon after the start of the trace when there is a fall of 1.6°F in 1,500 feet. There are also two other distinctive pulses. The first of these is of magnitude 1.0°F which is attained in 1,500 feet, and though somewhat distorted by the presence of smaller more rapid fluctuations, returns to its initial temperature in a further 2,000 feet. The second pulse is of comparable dimensions to the first. The third large variation of temperature is considerably more distorted than the two previous ones. It has a magnitude of 1.0°F which it attains in 2,000 feet, but the return of the temperature to its initial value is considerably influenced by smaller fluctuations, one of which has magnitude 0.4°F. At the end of the trace, it seems that a fourth pulse of temperature is commencing, since there is a rise of temperature of 0.9°F in 1,800 feet. The trace is noisy.

Hot wire anemometer.

Duration of trace:- 1 min. 55 sec., or 23,000 ft. airspace.
Sensitivity:- 8 kt. full scale.

The trace begins with a sinusoidal rise of two knots in 1,500 feet followed by an unsteady fall of six knots due to a change in airspeed. Thereafter there are numerous smaller fluctuations, rising to two to three knots in some cases, but becoming quiet towards the end. The trace is noisy.

/7.

7. Analysis of ground observations on smoke puffs.

(a) Mean wind.

Table 1

Height.	Mean wind.		Direction.
feet.	ft/sec.	kt.	degrees.
11,000	37.0	21.9	024.6
10,500	35.2	20.8	017.6
10,000	36.4	21.5	016.6
9,500	37.5	22.2	022.1
9,000	36.8	21.8	026.8

It is of interest to note in the above the backing of the wind from 024.6° at 11,000 feet through 17.6° at 10,500 feet to 16.6° at 10,000 feet, and then the veering to 26.8° at 9,000 feet.

(b) Horizontal Shear.

In order to calculate the mean value of the horizontal shear at each level in two directions at right angles to one another, we first evaluate the components of the wind given by the path of each smoke puff. Then knowing the time interval between successive puffs and the speed of the aircraft we can obtain the mean shear. The table below is obtained from the means for five pairs of successive puffs. We assume the X axis to be the 090 line, and the Y axis the 360 line.

Table 2

Height.	$\partial v/\partial x$	$\partial v/\partial y$
feet.	10^{-4} sec.$^{-1}$	10^{-4} sec.$^{-1}$
11,000	-1.86	-1.34
10,500	-0.36	-0.28
10,000	-1.98	-0.68
9,500	2.92	-3.35
9,000	0.60	0.89

(c) Vertical Shear.

Since the time interval between any two consecutive levels is of the same order as the time taken on each run, it seems advisable to obtain the mean vertical shear from the mean winds at each level. This has been done for the following table.

Table 3

Height range.	$\partial v/\partial z$
feet.	10^{-3} sec.$^{-1}$
11,000 - 10,500	9.5
10,500 - 10,000	-2.6
10,000 - 9,500	-7.4
9,500 - 9,000	6.3

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8. Evaluation of the Richardson number.

Defining the Richardson number as:-

$$Ri = \frac{g}{T} \frac{\left(\frac{\partial T}{\partial z} + \Gamma\right)}{\left(\frac{\partial v}{\partial z}\right)^2} ,$$

where T = mean temperature of layer in degrees F Absolute.

Γ = adiabatic lapse rate in degrees F per foot.

z = height in feet between the two levels considered.

v = vector wind velocity in ft/sec.

g = acceleration due to gravity in ft/sec^2,

the following values for 500 foot intervals between 9,000 feet and 11,000 feet were obtained.

Table 4

Height range. feet.	Ri.
9,000 - 9,500	3.0
9,500 - 10,000	1.9
10,000 - 10,500	7.9
10,500 - 11,000	0.7

It is convenient to arrange the data so obtained in a table which may be taken as a vertical plane drawn through the block of air under investigation (Table 5). It is at once clearly seen that the largest value of the Richardson number occurs in a layer at whose boundaries the temperature and wind fluctuations are least, whilst the smallest value of this number lies in a layer across which these fluctuations change from being very quiet at one edge to noisy at the other.

9. Conclusions.

(a) It seems that we now possess apparatus capable of measuring turbulence parameters in the free atmosphere. The main difficulties in this type of work from a practical point of view are the low frequency of bumpy conditions, and the necessity for clear ground-air visibility. Other limitations are the field of view of the camera obscura, and the low ceiling of the aircraft which can carry the necessary equipment.

(b) The first experiment showed some correlation between the value of the Richardson number and the amount of turbulence.

(c) The small fluctuations of temperature and wind are probably related and a desirable feature in future design is simultaneous recording. Work is proceeding on these lines.

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Height feet.	Mean Temp. °F	Mean Wind Direction (Degrees)	Mean Wind Speed Ft/Sec.	Mean F.P. °F	Ultra Rapid Thermometer.	Hot Wire Anemometer.	Horizontal Shear 10^{-4} $sec.^{-1}$	Vertical Shear 10^{-3} $sec.^{-1}$	Richardson Number.
11,000	20.1	024.6	37	13	Noisy 1-2°F variations	Noisy 2-3 kt gusts	-1.86 -1.34		
								9.5	0.7
10,500	22.3	017.6	35.2	14	Noisy 1-2°F variations	Quiet	-0.36 -0.28		
								-2.6	7.9
10,000	24.6	016.6	36.4	15	Quiet	Very quiet	-1.98 -0.68		
								-7.4	1.9
9,500	26.5	022.1	37.5	18	Noisy 1-2°F variations	Noisy 1-2 kt gusts	2.92 -3.35		
								6.3	3.0
9,000	28.3	026.8	36.8	21	Noisy 1-2°F variations	Noisy 2-4 kt gusts	0.60 0.89		

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At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 794 by Mr. F.H. Bushby & Miss Mavis K. Hinds. A copy is available in the Library of the Meteorological Office."

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25th March, 1953

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AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

Computation of the 1000 - 500 mb thickness tendency, the 1000 mb height tendency, and the horizontal field of vertical motion, using an electronic computer

by
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1. Introduction

Subject to various approximations, Sawyer and Bushby (1) have derived a set of simultaneous partial differential equations which, in theory, provide a method of forecasting the contour height of an isobaric surface representative of the mid-troposphere and the thickness of the layer from 1000 mb to that isobaric surface. Essentially the method consists of computing the instantaneous rates of change of the contour height and the thickness of the layer at time t, and hence computing the new contour and thickness patterns at time $t + \delta t$. By repeating the process n times, forecast charts can be prepared for time $t + n\delta t$.

The equation for the rate of change of height of the contour surface has been tested by application to three synoptic situations and the results have been reported by Bushby and Hinds (2). The present report describes the result of applying the thickness tendency equation to the same three situations: the horizontal field of vertical motion has also been computed. The 1000 mb height tendencies have been calculated by subtracting the computed thickness tendencies from the computed contour height tendencies, and they have been compared with the actual 1000 mb height tendencies.

2. The Method

In the model adopted, the state of the atmosphere at a given instant is specified by three parameters; h_m, the height of an isobaric surface $p = p_m$ near the middle of the troposphere; h', the thickness from a pressure level $p = p_o$ near the ground to $p = p_m$, and Γ_p, a stability parameter. Γ_p is a measure of the departure of the lapse rate from the appropriate wet or dry adiabatic defined by

$$\Gamma_p = \frac{\partial T}{\partial p} - \frac{\gamma}{gp} \qquad \text{.....(1)}$$

where γ is the appropriate wet or dry adiabatic lapse rate, ρ is density, g is the acceleration due to gravity and T is the absolute temperature. Since in the analysis of Sawyer and Bushby (1) Γ_p is assumed to be independent of pressure, it is consistent to measure a mean value of Γ_p through the layer under consideration. Thus the field of Γ_p can be obtained by a similar method to that described by Bushby (3). In the present work h_m has been identified with the height of the 500 mb contour surface, h' with the thickness of the layer from 1000 mb to 500 mb, and the horizontal field of Γ_p has been estimated for that layer.

Sawyer and Bushby give simultaneous differential equations for the thickness tendency and $\omega_m = dp/dt$ which is a measure of the vertical motion. Conformally representing the northern hemisphere by a stereographic projection from the south pole on to a plane perpendicular to the earth's axis and passing through the north pole, these equations are

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$$\nabla^2\frac{\partial h'}{\partial t} + J\left(h_m, \beta^2 g \ell^{-1}\nabla^2 h'\right) + J\left(h', \beta^2 g \ell^{-1}\nabla^2 h_m + \ell\right) = \frac{4\varpi_m\left(\nabla^2 h_m + \frac{\ell^2}{\beta^2 g}\right)}{p_0 - p_1}$$

$$\frac{\partial h'}{\partial t} = \beta^2 g \ell^{-1} J\left(h', h_m\right) - \frac{RAT_p \varpi_m}{g} \qquad \text{.....(3)}$$

where $$A \equiv \frac{1}{2}\frac{p_0 + 3p_1}{p_0 - p_1} - \frac{4p_0 p_1}{(p_0 - p_1)^2}\log_e \frac{2p_0}{p_0 + p_1}\ ; \qquad \text{.....(4)}$$

∇^2 is the two dimensional Laplacian operator referred to rectangular Cartesian co-ordinates; J is the Jacobian operator, β is the magnification factor $\sec^2\left(\frac{\pi}{4} - \frac{\theta}{2}\right)$, θ being the geographical latitude; ℓ is the Coriolis parameter; ϖ_m is the value of dp/dt at the level $p = p_m$; p_1 and p_0 are the pressures at the top and bottom of the troposphere, taken as 200 mb and 1000 mb in this case, and R is the universal gas constant.

Eliminating ϖ_m from Eqs. (2) and (3), we derive the following Helmholtz equation for $\partial h'/\partial t$:-

$$\nabla^2\frac{\partial h'}{\partial t} + \frac{4g\left(\nabla^2 h_m + \frac{\ell^2}{\beta^2 g}\right)}{RAT_p(p_0 - p_1)}\frac{\partial h'}{\partial t} + J\left(h_m, \beta^2 g \ell^{-1}\nabla^2 h'\right) +$$

$$+ J\left(h', \beta^2 g \ell^{-1}\nabla^2 h_m + \ell\right) + \frac{4g\beta^2 J\left(h_m, h'\right)}{RAT_p \ell (p_0 - p_1)} = 0 \qquad \text{.....(5)}$$

Computations based on Eqs. (5) and (3) were carried out on LEO, an electronic computing machine of similar type to the EDSAC, the Cambridge machine. A programme MET 5*, was made to solve Eq. (5) for $\partial h'/\partial t$ at a network of 12 X 8 points and then to solve for ϖ_m by substituting for $\partial h'/\partial t$ in Eq.(3).

The programme occupies 757 storage locations, whilst the data, intermediate and final results for an $r \times s$ grid would occupy $(8rs + 16r + 24s + 52)$ storage locations, 1204 for the 12 X 8 grid used. The total time taken by the programme was about 15 mins., of which 6 mins. was spent in reading programme and data, 4 mins. in printing results and 5 mins. in computing. The grid length, a, was about 160 miles in middle latitudes.

Eq. (5) was solved by a modified form of the Liebmann (4) iterative process. This process is more convergent when applied to a Helmholtz equation than when applied to a Poisson equation, and the results were printed after 30 iterations. The computations were carried out twice for each of the three situations that were examined. The actual values of T_p were used in the first set of computations, but in the second set a constant value of T_p (obtained by averaging the measured values of all 3 situations) was used. The results of the computations are shown in the appendix. The repetition of the computations, assuming T_p constant, has produced surprisingly little change in the computed values of $\partial h'/\partial t$ and ϖ_m , although the actual values of T_p varied considerably.

The 1000 mb height tendency, $\partial h_0/\partial t$, was calculated by hand from the following relationship

$$\frac{\partial h_0}{\partial t} = \frac{\partial h_m}{\partial t} - \frac{\partial h'}{\partial t} \qquad \text{.....(6)}$$

Values of $\partial h_m/\partial t$ had already been computed by Bushby and Hinds (2) using a similar electronic technique.

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In order to solve Eq. (5) it was necessary to make some assumption regarding the boundary of the area studied. For this purpose it was assumed that $\partial h'/\partial t = 0$ at the 44 points exterior to the 12 X 8 grid at which values were computed. In order to estimate the effect of errors arising due to the use of wrong boundary values, the correct values of $\partial h'/\partial t$ and $\partial h_o/\partial t$ on the boundary were estimated from working charts and the necessary corrections to the computed values of $\partial h'/\partial t$ and $\partial h_o/\partial t$ were calculated by hand. The effect of these corrections is discussed in section 3.

3. Results

The three synoptic situations for which computations were made are shown in Figs. 1a, 2a and 3a. The actual 1000 - 500 mb thickness tendency fields, averaged over a 12 hour period centred at the times for which the computations were made, are shown in Figs. 1b, 2b, 3b, whilst the computed thickness tendency fields, assuming zero boundary values, and using a variable field of T_p, are shown in Figs. 1c, 2c, and 3c. The actual 1000 mb height tendency fields, averaged over a 6 hour period centred at the times for which the computations were made, were measured from the surface charts and are shown in Figs. 1d, 2d, 3d. The computed fields of $\partial h_o/\partial t$, assuming zero boundary values, are shown in Figs. 1e, 2e, and 3e. It can be seen from the diagrams that there is a reasonable agreement between the computed and actual tendency fields in all three situations.

Correlation coefficients have been calculated comparing the computed tendencies with the actual tendencies, and regression coefficients were calculated for the best fitting straight line giving the actual rate of change (Y feet per hour) in terms of the computed rate of change (X feet per hour) in the form

$$Y = bX + c \qquad \text{.....(7)}$$

and the results are shown in Table I. The R.M.S. differences between the actual and computed tendencies are given in Table II. The notation used in Tables I and II is as follows:-

A ≡ Actual $\partial h'/\partial t$ (ft./hr.)

B ≡ Computed " (ft./hr.), zero boundary values.

C ≡ Computed " (ft./hr.), actual boundary values.

D ≡ Actual $\partial h_o/\partial t$ (ft./hr.)

E ≡ Computed " (ft./hr.), zero boundary values.

F ≡ Computed " (ft./hr.), actual boundary values.

Table I

Correlation and regression coefficients of 1000 - 500 mb thickness tendencies and 1000 mb height tendencies

		1500 G.M.T. 14.3.49			1500 G.M.T. 27.1.52			1500 G.M.T. 20.12.51		
Y	X	r	b	c	r	b	c	r	b	c
A	B	0.49	0.39	-0.4	0.76	0.67	1.5	0.85	0.94	2.4
A	C	0.72	0.49	0.2	0.80	0.61	-0.1	0.90	0.80	2.2
D	E	0.93	0.98	-0.3	0.67	0.72	-0.9	0.83	1.18	-8.0
D	F	0.95	0.68	-0.1	0.78	0.73	2.7	0.88	1.07	-1.3

r = correlation coefficient.

/Table II

Table II

Root mean squares in ft./hr.

	1500 G.M.T. 14. 3.49.	1500 G.M.T. 27. 1.52.	1500 G.M.T. 20.12.51.
A	10.5	10.6	27.3
A - B	12.3	8.9	13.7
A - C	10.8	8.3	12.7
D	17.1	12.0	22.7
D - E	5.9	9.2	14.6
D - F	9.2	9.4	10.7

It is evident from Table I that there is significant correlation between the computed and actual tendency fields, this being greater when the correct boundary values were applied than when zero boundary values were used. These results are confirmed by Table II for two of the situations, but for 1500 G.M.T. on the 14th March, 1949, the R.M.S. of (A - B) is greater than that of A and the R.M.S. of (D - E) is less than that of (D - F), which may at first seem inconsistent with the correlation coefficients of Table I. However, this particular situation was one in which a rapidly moving wave depression, followed by a mobile ridge, was moving E.S.E. across the British Isles and Eastern Atlantic and one would therefore not expect the computed instantaneous rates of change to be of the same magnitude as the measured rates of change meaned over a six or twelve hour period, although the tendency patterns should correspond. This point emphasises the fact that it is difficult to obtain a satisfactory objective method of comparing two sets of isopleths of this type.

The field of vertical motion computed for 1500 G.M.T. 14th March, 1949, is shown in Fig. 1f and indicates an area of ascending motion ahead of the mobile wave depression, with a large region of descending motion ahead of the ridge over the eastern Atlantic. Another area of ascending motion is shown ahead of the central Atlantic troughs. In this situation the computed vertical motion agrees quite well with that expected from synoptic experience.

Fig. 2f shows the vertical motion field that was computed for 1500 G.M.T. 27th January, 1952, and the main feature is the extensive region of subsidence indicated over southern France and the Bay of Biscay. This was associated with a surface pressure ridge and is on the western side of a cold trough. Other features of interest are the areas of convergence ahead of the frontal troughs in the eastern Atlantic and southern Scandinavia.

The vertical motion field computed for 1500 G.M.T. 20th December, 1951, is shown in Fig. 3f. This is an extremely interesting case as the twin centres of subsidence to the east of Scotland and to the south-west of Ireland, with an area of convergence over western France, are symptomatic of the cut-off process and anticyclonic building across the neck of a cold trough. This development actually occurred.

4. Conclusions

A rather unexpected result of the present work was that the computed 1000 mb height tendencies, obtained by subtracting the thickness tendencies from the contour height tendencies, should show such good agreement with actuality. As the thickness and the contour tendencies are of the same order of magnitude, it was expected that any real difference between the two quantities would be swamped by errors arising from the basic assumptions and the approximate methods of the computations. This does not appear to

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be true, at least for the three situations so far examined, and it is possible that the present model may give a satisfactory indication of surface pressure changes.

The main conclusion is that the Sawyer - Bushby model does give a good representation of the 1000 - 500 mb thickness changes and of the fields of vertical motion in the three synoptic situations to which it has been applied, even if the field of static stability is assumed constant. As a similar degree of success was achieved when the 500 mb height tendency equation was tested, the present results appear to justify an attempt to produce prognostic charts by the simultaneous time integration of the contour and thickness tendency equations.

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$\bar{w}_m$ in mb./hr. (T_p variable) for 1500 G.M.T. 14.3.49.

-22	-16	-6	8	27	15	-8	1	10	5	4	6
-15	-13	-1	8	11	20	14	-2	2	9	5	3
-15	-11	1	9	11	13	8	-2	-5	1	2	1
-7	-1	2	7	14	14	3	-19	-23	-3	6	4
-2	0	4	6	5	2	-4	-13	-9	1	3	1
-3	-1	2	1	1	-2	-3	-3	0	2	1	1
-1	-1	0	1	1	0	-2	0	3	2	1	0
3	2	0	1	0	-1	0	2	2	0	0	0

Table IID

$\bar{w}_m$ in mb./hr. (T_p constant) for 1500 G.M.T. 14.3.49.

-24	-17	-7	10	27	13	-4	1	7	4	4	5
-15	-13	-2	9	12	20	13	-2	2	7	4	3
-14	-12	0	9	11	14	8	-2	-3	1	2	1
-6	-1	1	8	15	15	3	-19	-21	-3	8	5
-2	0	3	7	4	3	-6	-10	-5	0	4	1
-3	-1	1	2	1	-2	-5	-3	00	1	2	0
-2	0	0	1	0	-1	-1	-1	2	2	1	0
3	2	0	2	0	-1	0	2	2	0	0	0

/Table IIA.

Table IIA

$\partial h'/\partial t$ in ft./hr. (T_p variable) for 1500 G.M.T. 27.1.52.

-13	-12	-6	-1	2	5	2	-2	3	8	3	0
-16	-18	-10	-3	2	4	3	-4	-2	0	-4	0
-15	-14	-12	-9	-2	3	4	-2	2	-2	-7	-8
-14	-15	-18	-16	-6	8	9	9	7	-5	-3	-9
-15	-27	-32	-19	4	13	16	18	10	-14	-14	-9
-21	-27	-25	-13	12	18	21	21	6	-19	-24	-13
-10	-19	-17	2	20	17	12	11	3	-10	-21	-8
-2	-5	-1	7	14	14	6	2	-1	-4	-19	-9

Table IIB

$\partial h'/\partial t$ in ft./hr. (T_p constant) for 1500 G.M.T. 27.1.52.

-13	-11	-6	0	1	5	3	-2	3	8	2	0
-16	-18	-9	-3	1	5	4	-4	-2	-2	-5	0
-15	-13	-12	-10	-1	3	4	-2	2	-3	-8	-9
-13	-14	-18	-18	-6	8	9	9	7	-5	-2	-9
-15	-27	-32	-20	3	12	16	18	11	-14	-13	-9
-21	-27	-25	-13	11	17	20	21	8	-18	-25	-13
-10	-18	-18	2	21	16	11	11	3	-9	-21	-8
-2	-5	-1	7	14	13	6	2	-1	-4	-19	-9

Table IIC

ϖ_m in mb./hr. (T_p variable) for 1500 G.M.T. 27.1.52.

4	5	-1	-5	-2	3	4	-1	-3	0	2	-1
7	4	0	-4	-5	0	4	2	-1	-6	-5	1
8	6	0	-6	-8	-2	3	0	-2	-2	-9	-5
4	9	5	-9	-11	0	1	3	2	4	-1	-8
4	-1	-3	-8	-8	-1	4	1	3	0	2	-9
1	-3	-5	-11	-7	1	8	4	9	0	-10	-12
1	-8	-9	-6	4	8	7	8	15	17	-2	1
-3	-5	-4	-2	7	10	7	8	12	21	-1	-12

Table IID

ϖ_m in mb./hr. (T_p constant) for 1500 G.M.T. 27.1.52.

3	5	0	-4	-2	3	4	0	-3	0	1	-1
7	3	1	-4	-6	1	4	1	0	-5	-5	1
7	6	0	-6	-7	-2	4	0	-1	-2	-7	-5
4	7	4	-8	-12	0	2	3	1	3	0	-8
4	0	-2	-7	-8	-2	4	1	3	0	. 2	-9
1	-3	-5	-10	-7	0	7	4	8	0	-9	-13
1	-8	-11	-7	5	8	7	8	15	19	-2	1
-3	-6	-4	-3	8	11	8	7	12	21	-1	-11

/Table IIIA.

Table IIIA.

$\partial h'/\partial t$ in ft./hr. (T_p variable) for 1500 G.M.T. 20.12.51.

-25	18	60	62	29	2	-20	-44	-26	-1	6	5
-28	35	77	67	26	6	-17	-32	-37	-7	4	5
-11	48	77	53	18	5	-17	-21	-27	-14	-2	1
2	37	50	36	14	5	-12	-15	-22	-18	-9	-3
14	28	36	36	32	16	-2	-8	-14	-11	-5	-2
17	21	31	32	29	6	-1	-16	-7	2	1	1
5	7	4	0	6	-7	-10	-10	-4	7	4	3
1	1	-2	8	2	-7	-7	-1	1	0	1	2

Table IIIB.

$\partial h'/\partial t$ in ft./hr. (T_p constant) for 1500 G.M.T. 20.12.51

-25	18	60	62	30	3	-18	-41	-25	-1	7	6
-29	35	77	66	26	7	-15	-30	-35	-6	5	6
-13	48	76	53	18	5	-17	-21	-26	-13	-1	2
1	37	50	36	14	6	-12	-15	-22	-17	-8	-2
13	29	36	35	32	17	-2	-10	-14	-11	-5	-2
18	21	31	30	28	6	-2	-17	-8	2	1	1
5	8	5	-1	4	-7	-10	-12	-5	8	4	2
2	2	-2	7	0	-7	-7	-3	1	0	2	1

Table IIIC.

ω_m in mb./hr. (T_p variable) for 1500 G.M.T. 20.12.51.

19	-21	-15	12	5	1	10	7	-1	-5	-6	-4
8	-4	4	9	1	0	7	18	-1	-4	-2	-2
9	0	3	0	-5	4	-2	5	-4	-4	-3	-3
4	-1	-5	-8	-6	6	0	0	-10	-6	-3	-2
-2	-2	4	8	17	11	1	-2	-9	-2	-1	0
-6	-3	5	11	16	4	-5	-12	-2	2	1	1
-9	-7	-9	1	8	4	-5	-9	-3	1	2	3
-5	-4	-5	10	11	5	-3	-6	-8	-4	-2	3

Table IIID.

ω_m in mb./hr. (T_p constant) for 1500 G.M.T. 20.12.51.

18	-22	-17	13	5	1	7	8	-1	-7	-7	-5
9	-4	5	9	1	1	6	17	0	-5	-3	-3
12	0	3	0	-5	3	-2	5	-4	-5	-3	-4
5	-1	-6	-9	-6	5	0	0	-10	-6	-3	-2
-4	-2	4	8	17	11	1	-4	-7	-3	-1	0
-7	-4	6	11	17	4	-5	-11	-2	2	2	2
-12	-8	-10	0	7	4	-5	-8	-3	2	3	3
-6	-4	-6	11	11	5	-2	-5	-7	-5	-1	2

M.R.P. 794

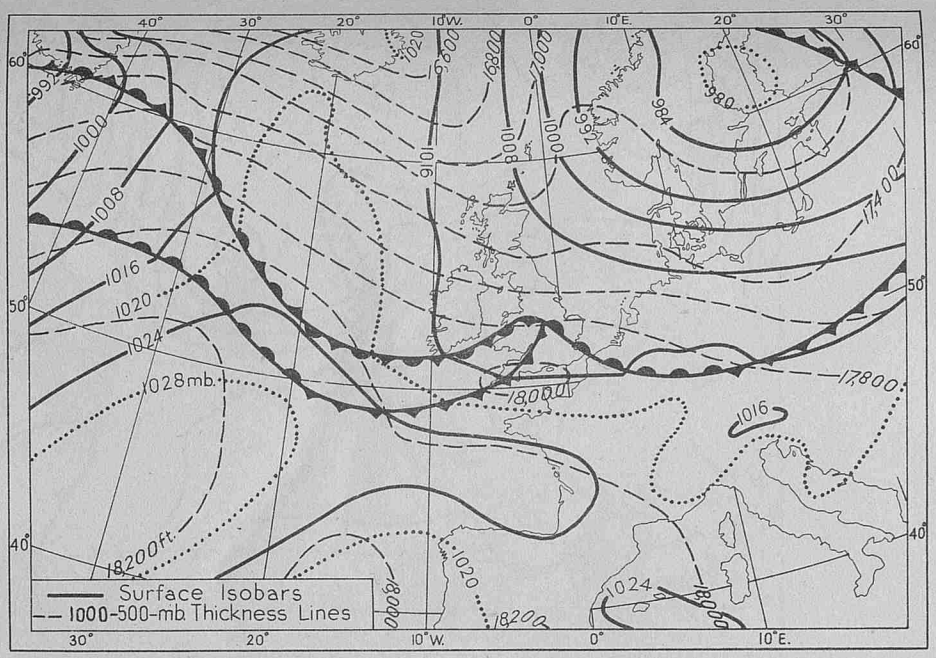

Fig. 1a SYNOPTIC CHART 1500 G.M.T. 14.3.49

M.R.P. 794

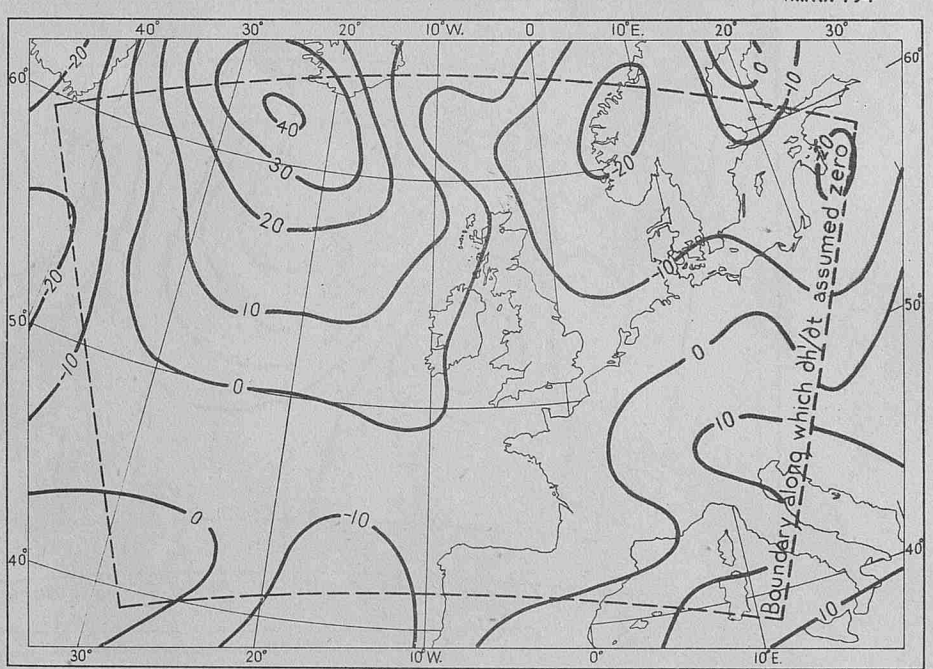

Fig. 1b ACTUAL 1000-500-mb. THICKNESS TENDENCY (ft./hr.) 1500 G.M.T. 14.3.49

M.R.P. 794

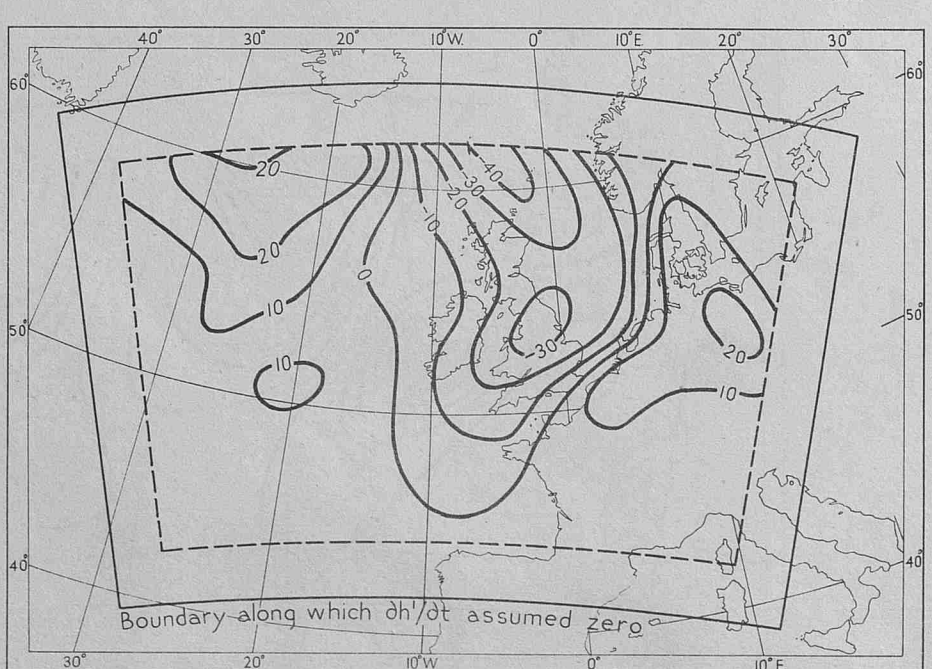

Fig. 1c COMPUTED 1000-500-mb. THICKNESS TENDENCY (ft./hr.) 1500 G.M.T. 14.3.49

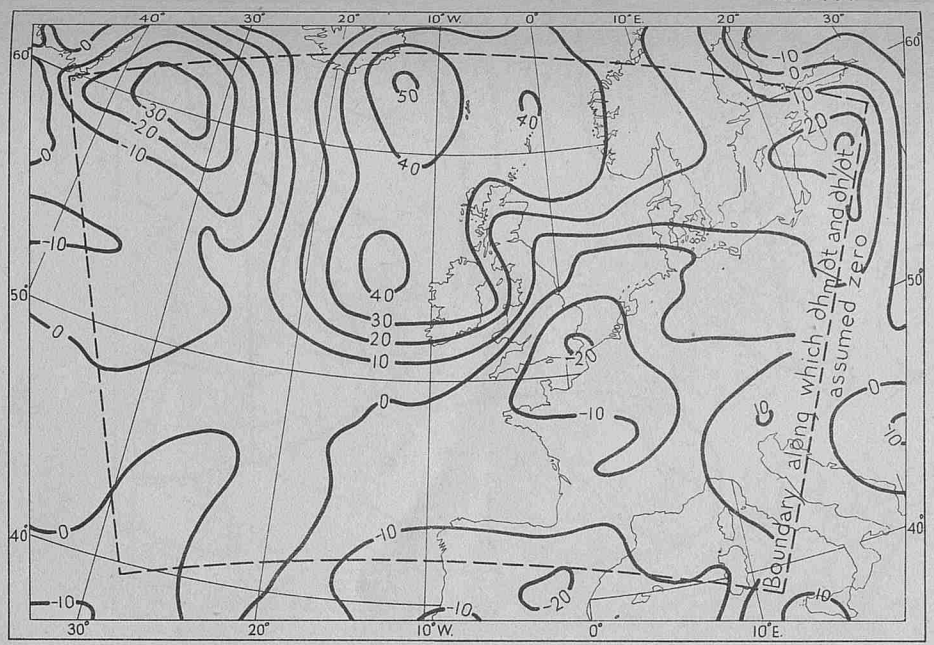

Fig. 1d ACTUAL 1000-mb. HEIGHT TENDENCY (ft./hr.) 1500 G.M.T. 14. 3. 49

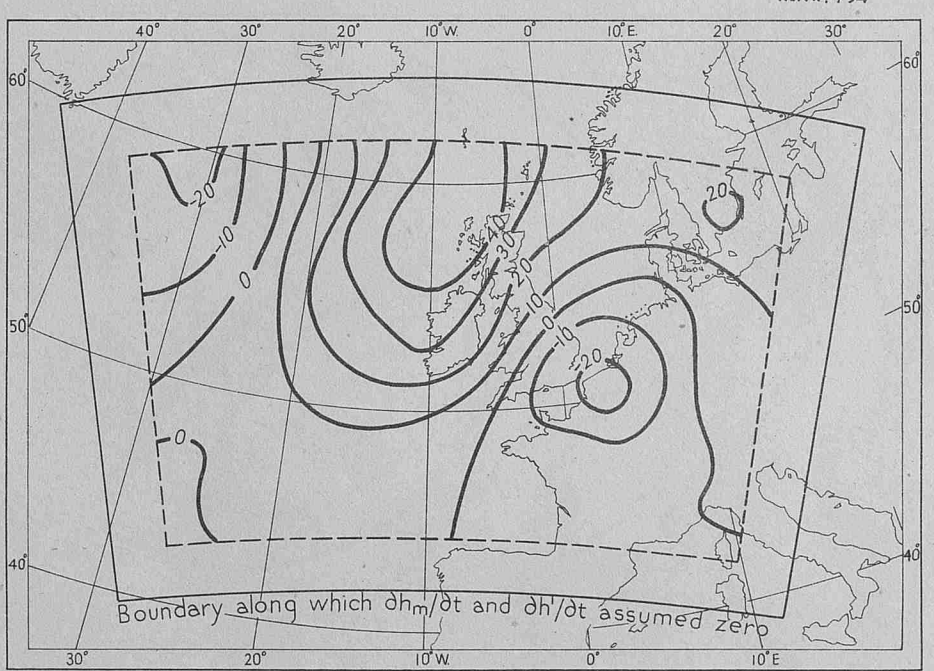

Fig. 1e COMPUTED 1000-mb. HEIGHT TENDENCY (ft./hr.) 1500 G.M.T. 14. 3. 49

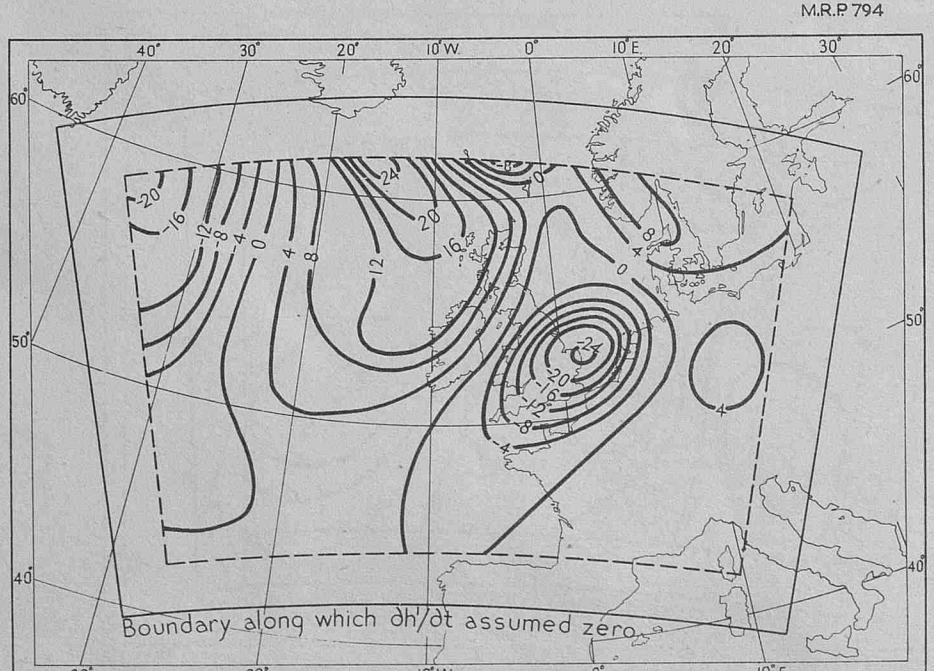

Fig. 1f VERTICAL MOTION (dp/dt in mb/hr.) 1500 G.M.T. 14. 3. 49

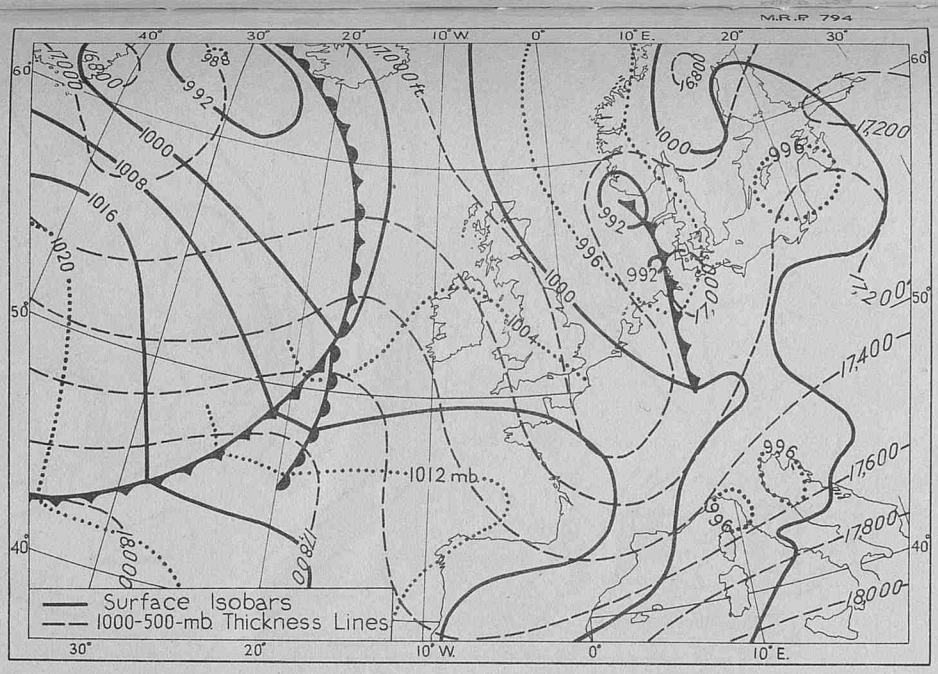

Fig. 2a SYNOPTIC CHART 1500 G.M.T. 27. 1. 52

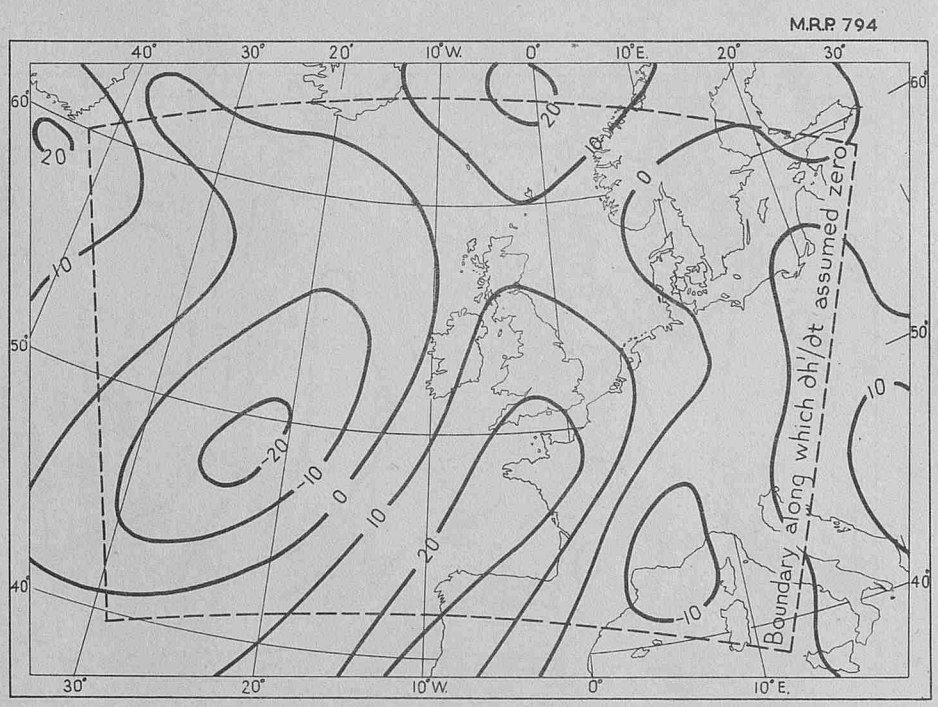

Fig. 2b ACTUAL 1000–500-mb. THICKNESS TENDENCY (ft/hr.) 1500 G.M.T. 27. 1. 52

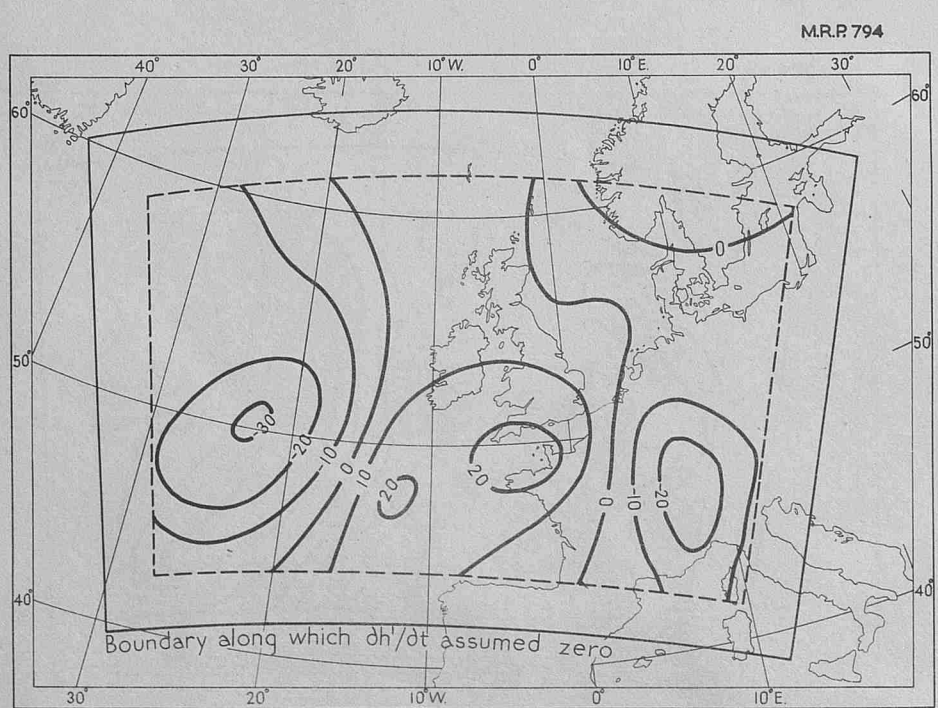

Fig. 2c COMPUTED 1000–500-mb. THICKNESS TENDENCY (ft/hr.) 1500 G.M.T. 27. 1. 52

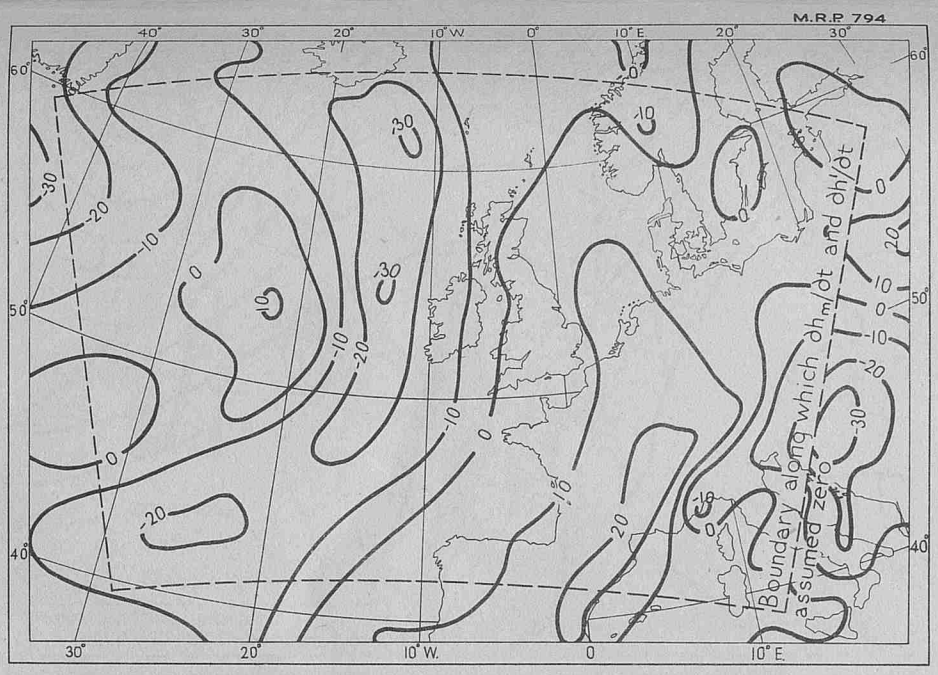

Fig. 2d ACTUAL 1000-mb. HEIGHT TENDENCY (ft./hr.) 1500 G.M.T. 27. 1. 52

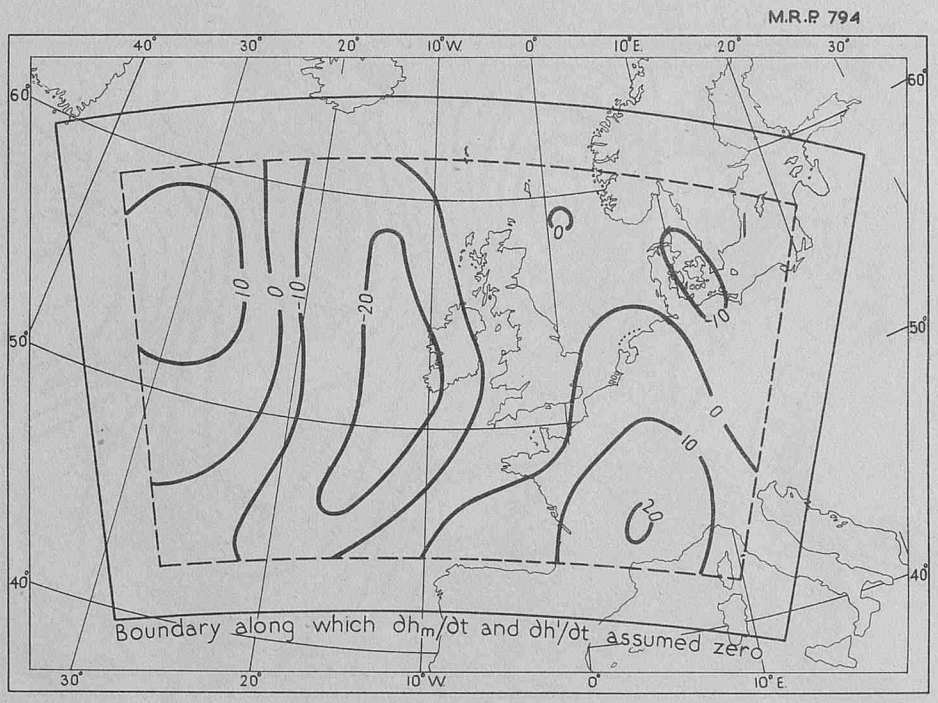

Fig. 2e COMPUTED 1000-mb. HEIGHT TENDENCY (ft/hr.) 1500 G.M.T. 27. 1. 52

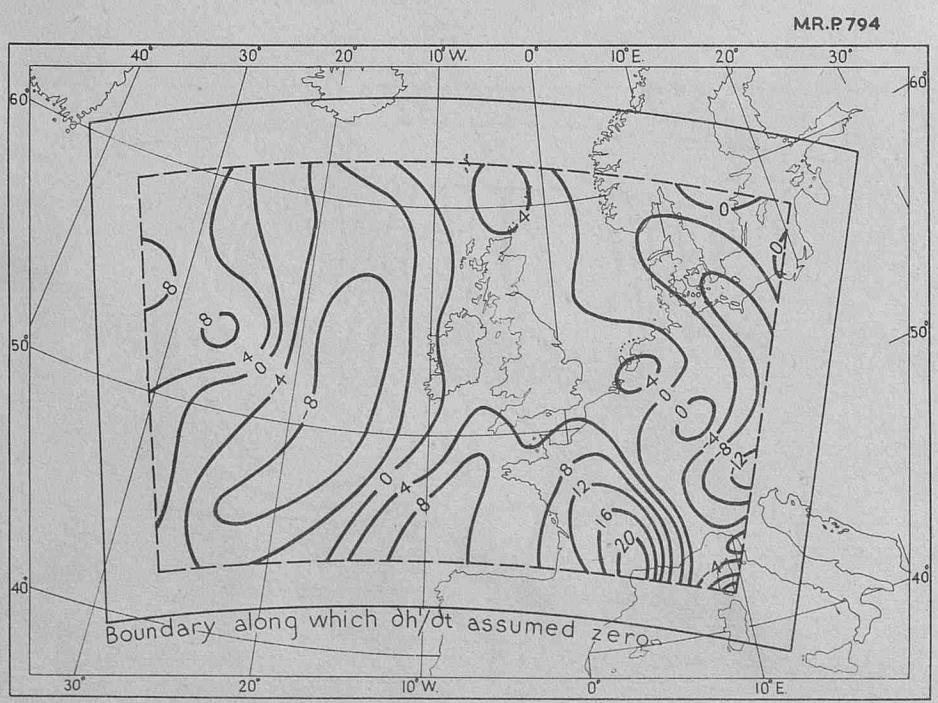

Fig. 2f VERTICAL MOTION (dp/dt in mb./hr.) 1500 G.M.T. 27.1 52

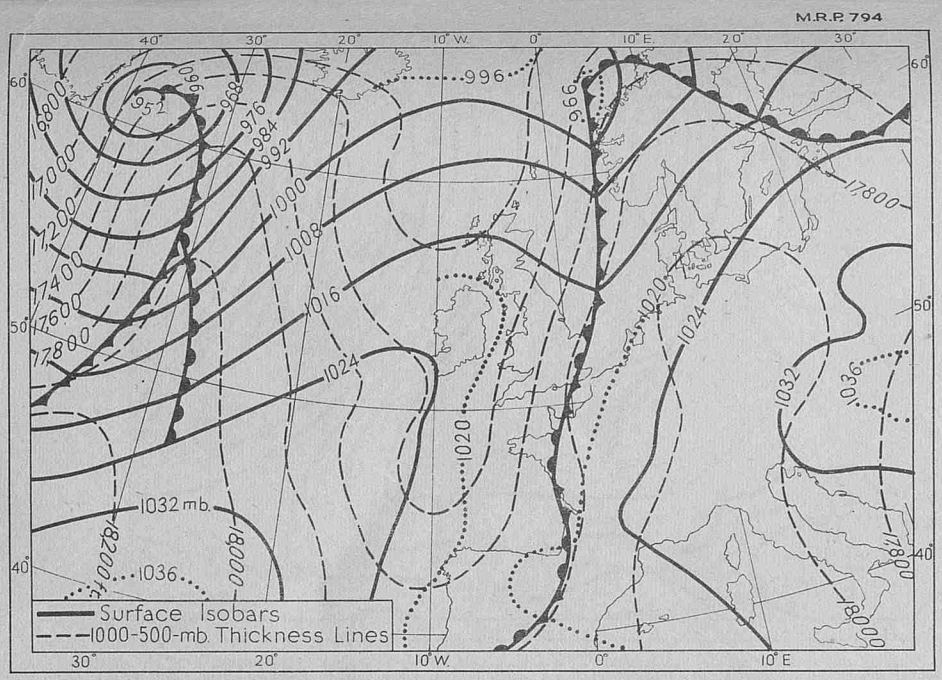

Fig. 3a SYNOPTIC CHART 1500 G.M.T. 20.12.51

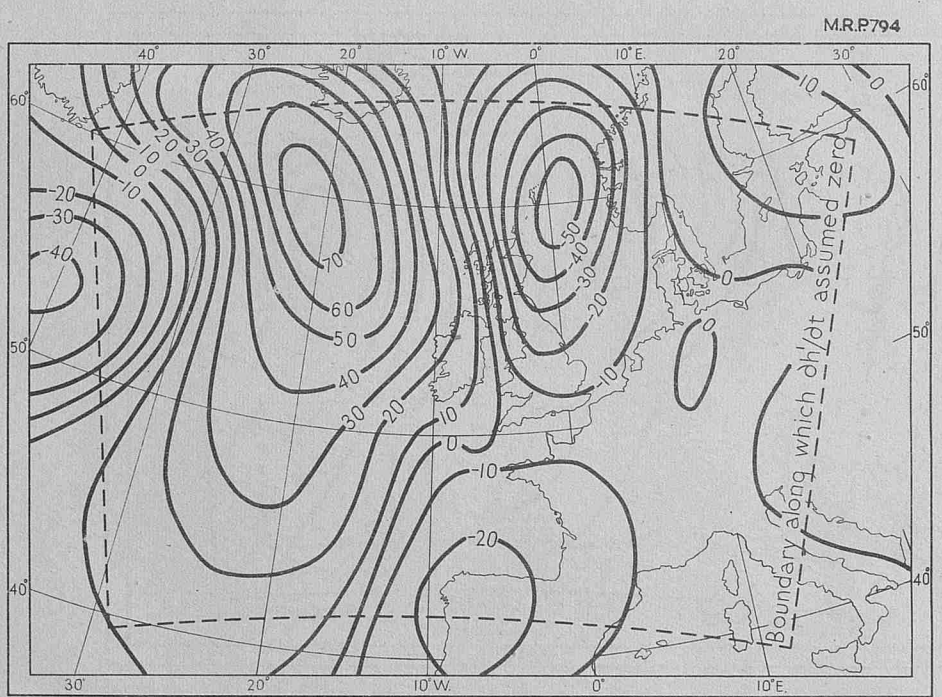

Fig. 3b ACTUAL 1000-500-mb. THICKNESS TENDENCY (ft./hr.) 1500 G.M.T. 20.12.51

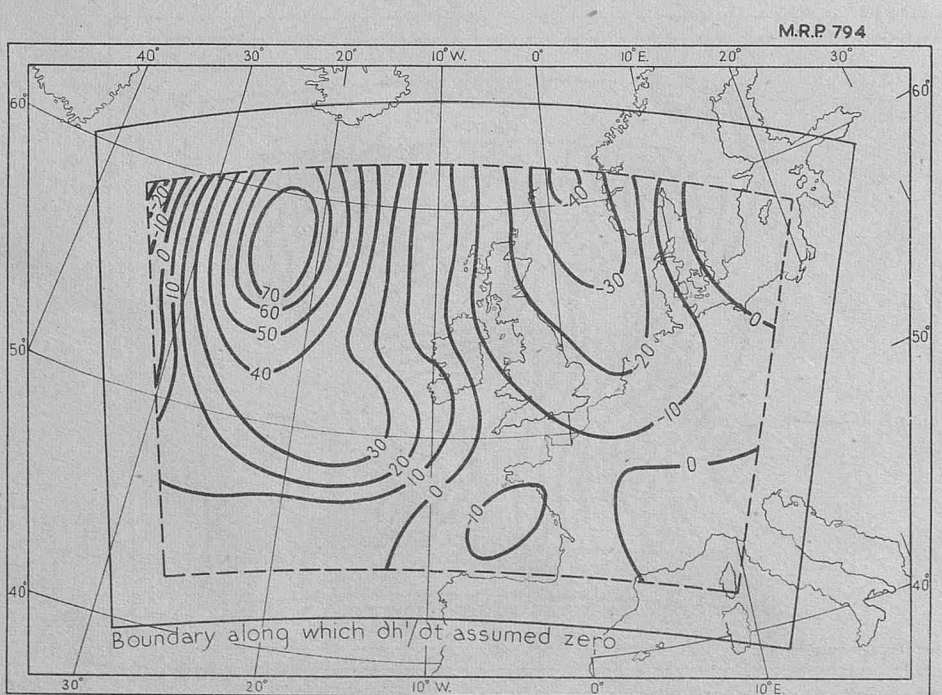

Fig. 3c COMPUTED 1000-500-mb. THICKNESS TENDENCY (ft./hr.) 1500 G.M.T. 20.12.51

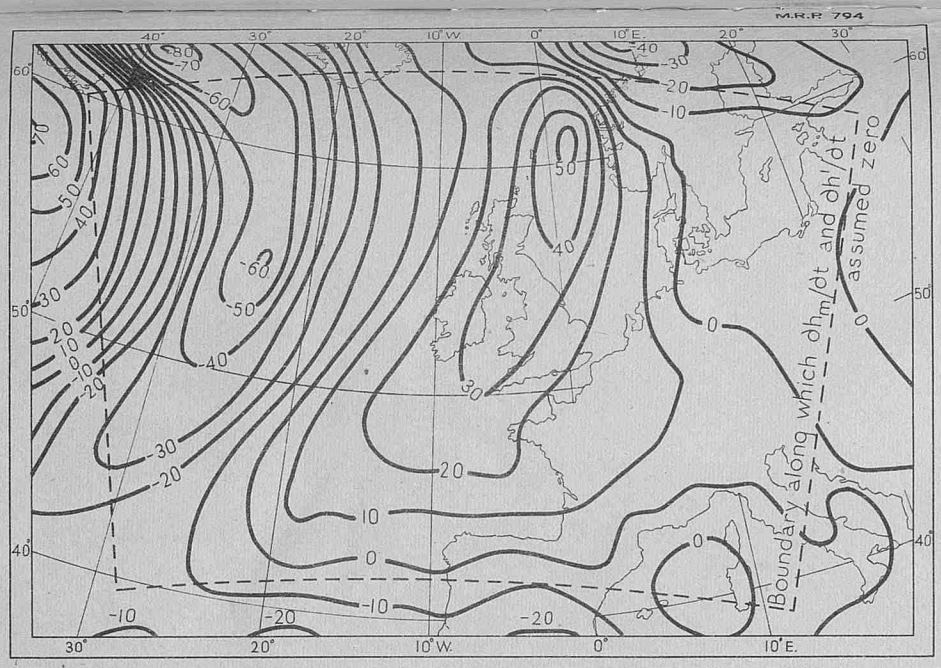

Fig. 3d ACTUAL 1000-mb. HEIGHT TENDENCY (ft./hr.) 1500 G.M.T. 20.12.51

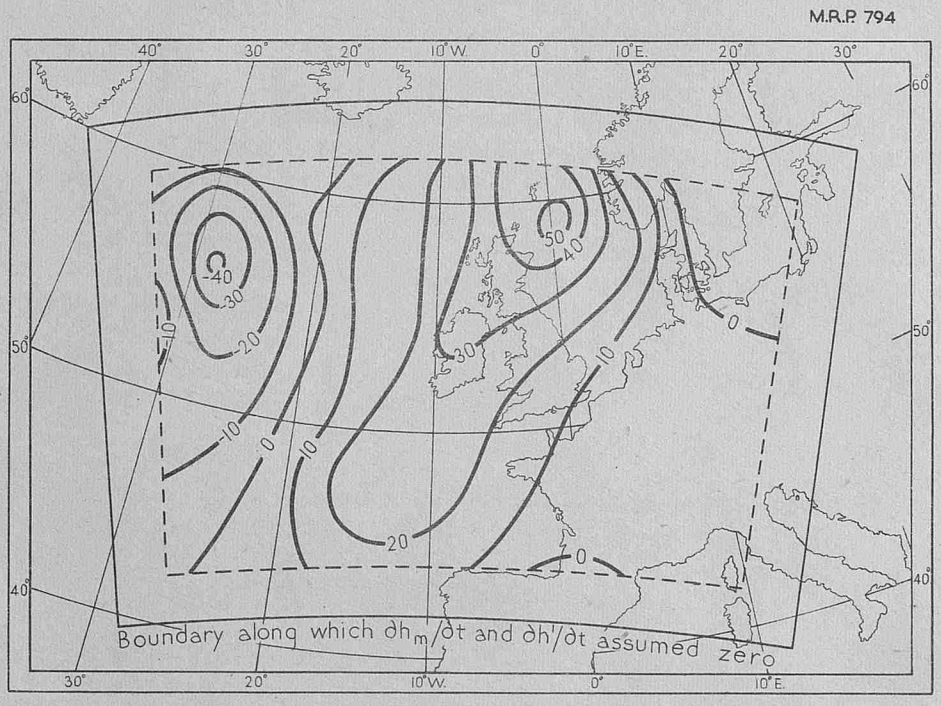

Fig. 3e COMPUTED 1000-mb. HEIGHT TENDENCY (ft./hr.) 1500 G.M.T. 20.12.51

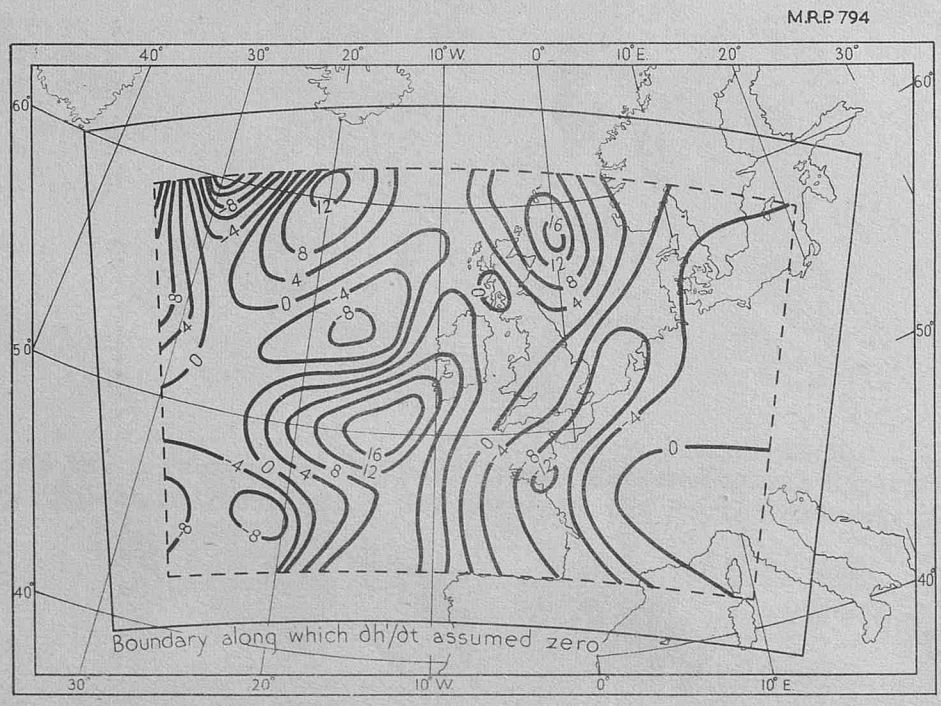

Fig. 3f VERTICAL MOTION (dp/dt in mb/hr.) 1500 G.M.T. 20.12.51

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 795 by Mr. A. Gilchrist

A copy is available in the Library of the Meteorological Office."

M.R.P. 795
S.G.II/141
10th April, 1953.

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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

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Upper Winds in the Tropics and Subtropics

by

A. Gilchrist (Meteorological Office)

INTRODUCTION

This investigation was undertaken to provide a rapid survey of upper winds in low latitudes, as a preliminary to the fuller and more comprehensive study which will be necessary for the revision of Geophysical Memoir No. 85, "Upper Winds over the World". Attention has been concentrated on the four mid-season months of 1951, for which all obtainable observations of high level winds from tropical and subtropical regions were collected. By means of charts showing mean winds at constant pressure levels and mean atmospheric cross-sections at selected longitudes an attempt has been made to present an account of the mean atmospheric flow in the months considered. Cross-sections at 80°W. for particular days showing the structure of the wind maximum to the south of the polar jet-stream have also been drawn.

(Note. This paper was prepared while the author was working in the Upper Air Climatology Branch (M.O.23) during part of his training as a Scientific Officer.

The work was done under the guidance of Mr. J.K. Bannon. Assistance in the computations was given by the staff of the Branch.)

INFORMATION PRESENTED

1. MEAN WIND CHARTS

Charts showing the mean winds for January, April, July and October at 300 mb., 200 mb., 150 mb. and 100 mb. have been prepared (Figs. 1-4). These winds were calculated from one month's observations for as many stations as possible between 40°N. and 40°S. As far as possible, the data used were for 1951; but where observations from a station were not available, those for another year were substituted. The years of those observations which were not for 1951 are given in the Appendix where the names and positions of all stations used have been listed.

The sources of information were:-

M/S records in the Meteorological Office including observations for Singapore (1);
U.S. National Climatological Summary - U.S. Weather Bureau;
Daily Series Synoptic Weather Maps - U.S. Weather Bureau;
Aerological Data of Japan;
New Zealand Synoptic Observations;
Indian Daily Weather Reports;
Unpublished data made available through the courtesy of the Chief of the U.S. Weather Bureau and the Directors of the East Africa Meteorological Department, the British West African Meteorological Services and the Royal Observatory, Hong Kong.

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The winds were calculated directly from the available data, no attempt being made to correct for losses. They are therefore subject to error due either to (1) losses of balloons in high winds or to (2) the irregularity or infrequency of the observations.

Losses of balloons in high winds affect particularly measurements made in the strong Westerly wind belt, which is found between 30°-50° North and South of the Equator, and means calculated from such measurements will be biased towards low values. Also, since the errors depend upon the performance of the wind-determining apparatus employed, they may be different for different countries. Thus, the measurements made at British-controlled stations in 1951 are probably more reliable than those of other countries because of the greater range of the [illegible]r apparatus (and possibly, to some extent, because the records available a[illegible]ore complete). It seems likely that even in the regions of strong wind the directions as plotted are reliable because the day to day changes of direction within the strong wind belt are small; the speeds may be corrected where necessary from the mean cross-section.

Where it appeared that the calculated mean winds had been rendered unrepresentative because the observations were too few or too irregular, no value was entered on the charts.

2. MEAN CROSS-SECTIONS

Cross-sections showing isopleths of mean zonal wind and potential temperature were drawn for three longitudes 80°W., 45°E., 140°E., with a supplementary one at 165°E., for the months January, April, July and October, 1951 (Figs. 5-20.)

In equatorial regions observed winds only were used, but North of about 25°N. these were supplemented where necessary by geostrophic winds, derived from the temperature field. This was done by determining thermal winds from the isopleths of potential temperature as described by Matthewman (2), and these were then added to the observed winds at a lower level to give the high level values; the observed mean winds at 700 mb. were accepted as reliable, and geostrophically derived corrections were applied only above this level.

Various authors (3)(4)(5)(6)(7) have suggested that use of the geostrophic assumption gives an overestimate of the true wind. Generally it has been the experience during this investigation that mean geostrophic winds, derived from the contour heights, do give values in excess of the measured winds, and it has not appeared possible to account for the excess by the curvature of the mean contours. It is considered, however, that the procedure adopted of basing the cross-sections largely upon observed winds and of using thermal winds as far as possible merely to correct these when they were inaccurate has minimised possible errors from this source. As a result some of the winds shown are noticeably less than those which would have been found from the mean contours. Thus, in the cross-section for which most reliance had to be placed on the geostrophic assumption, namely that for January 1951 at 140°E. (Fig. 8), the maximum wind speed shown (the last isopleth is 180 knots) is less than the value obtained geostrophically in a Japanese investigation (5) (210 knots approx.). The fact that five measurements at 12 km. give a mean speed of 150 knots at Tateno, supports the belief that the value of the maximum has not been overestimated.

The vertical lines drawn from the stations in the cross-sections indicate the height to which wind observations were available. Temperature soundings were always available to much greater heights.

/(a)

(a) 80°W - Caribbean Cross-Section. (Figs. 5, 9, 13, 17). The stations used in constructing these cross-sections were:-

(A.F.)	Albrook Field	08°59'N.	79°34'W.	R/W and R/S.
(S.I.)	Swan Island	17°24'N.	83°56'W.	R/S.
(S.J.)	San Juan	18°28'N.	66°06'W.	R/W
(Ca)	Camaguey	21°24'N.	77°55'W.	R/S.
(Ha)	Havana	23°09'N.	82°21'W.	R/W and R/S.
(Mi)	Miami	25°49'N.	80°17'W.	R/W and R/S.
(Ta)	Tampa	27°58'N.	82°32'W.	R/S.
(Bu)	Burrwood	28°59'N.	89°22'W.	R/W
(Ch)	Charleston	32°54'N.	80°02'W.	R/W and R/S.
(Gr)	Greensboro	36°05'N.	79°57'W.	R/W and R/S.
(Wa)	Washington	38°51'N.	77°02'W.	R/W and R/S.
(Pi)	Pittsburgh	40°21'N.	79°56'W.	R/S.

Of these stations, Burrwood and San Juan are considerably West and East respectively of the line of the cross-section; their use was dictated by the absence of R/W stations at their latitudes nearer 80°W. This is not a serious consideration when the flow is approximately zonal, but it may produce misleading results when it is not. However, it has always been possible to estimate the relevance of their observations from the temperature field, and, when necessary, to adjust the wind speed isopleths accordingly.

In these cross-sections the tropopauses were determined by extrapolating the mean temperature curves to find the height of minimum temperature. This method has been used by others (3), (4), (8) in the construction of mean cross-sections, but it is realised that the value obtained is not necessarily that which would be found by averaging the daily tropopause heights (4). Especially, near the region of overlap of the polar and tropical tropopauses the method may produce unsatisfactory results. The cross-sections, therefore, should be regarded as giving merely the general level at which the tropopauses occurred. It may be remarked, however, that the seasonal variation of the tropical tropopause, high in January and April, lower in July and October, seems to be now well substantiated. Hess (8) has shown that at 80°W the tropical tropopause is higher in January - February than in July - August. Similar, though less pronounced, changes are found in the Middle East (Figs. 6, 10, 14, 18).

(b) 45°E - Middle East Cross-Sections. (Figs. 6, 10, 14, 18). The stations used in these cross-sections were:-

(Na)	Nairobi	01°17'S.	36°50'E.	R/W and R/S.
(Ad)	Aden	12°50'N.	45°01'E.	R/W " R/S.
(Ba)	Bahrein	26°16'N.	50°37'E.	R/W " R/S.
(Ha)	Habbaniya	33°22'N.	43°24'E.	R/W " R/S.

For some months, winds and temperatures at 700 and 500 mb. were available from the Russian stations.

(Er)	Erevan	40°08'N.	44°28'E.
	Astrakhan	46°25'N.	48°03'E.

At the Northern end, the cross-sections at higher altitudes, and in some at all altitudes were completed by making use of long term mean temperatures from other Russian stations.

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In April 1951, only four ascents were made from Nairobi. For that month the mean winds and temperatures for the period 1948-50 were used. The following table gives the mean components for the three years separately.

	700 mb.		500 mb.		300 mb.		200 mb.		150 mb.		100 mb.	
	N. E. (knots)	Obs.	N. E. (knots)	Obs.	N. E. (knots)	Obs.	N. E. (knots)	Obs.	N. E. (knots)	Obs.	N. E. (knots)	Obs.
1948	3 11	34	0 13	34	1 5	34	-3 -3	34	-3 -7	33	0 4	29
1949	4 7	22	3 12	22	1 11	21	3 7	21	0 2	21	0 3	20
1950	5 13	9	2 10	9	1 3	9	0 0	9	-3 -8	9	3 -7	9

The variation from year-to-year is sufficient to change the cross-section only in minor detail. It seems likely therefore that the substitution of 1948-50 for 1951 does not impair the general validity of the cross-section.

The tropopauses have been determined using the definitions stated in M.O.555 Upper Air Data, Part 3 Habbaniya.

(c) 140°E. - Cross-Section through Japan (Figs. 8, 12, 16, 20). The stations used in this cross-section were:-

(Ya)	Yap	09°24'N.	138°08'E.	P.B. and R/S.
(Gu)	Guam	13°27'N.	144°44'E.	R/W " R/S.
(I.J.)	Iwo Jima	24°47'N.	141°20'E.	R/W " R/S.
(Tare)	O.W.S. Tare	29°N.	135°E.	R/W " R/S.
(Sh)	Shionomisaki	33°27'N.	135°46'E.	R/W " R/S.
(Ta)	Tateno	36°03'N.	140°08'E.	R/W " R/S.
(Wa)	Wajima	37°23'N.	136°54'E.	R/W " R/S.
(Se)	Sendai	36°16'N.	140°54'E.	R/W " R/S.
(Ak)	Akita	39°43'N.	140°06'E.	R/W " R/S.
(Sa)	Sapporo	43°04'N.	141°20'E.	R/W " R/S.

Yap gave no winds for July, and was not available for October. Iwo Jima provided winds to only 500 mb. in July, and was not available for October. To complete this part of the cross-section reference was made to the observations from Okinawa (26°12'N 127°45'E) and Marcus Island (24°17'N 153°58'E) and to the 1950 observations for Iwo Jima.

The tropopauses shown are based mainly on values of tropopause pressure given by the Japanese; the method of determination is not known. Experience of dealing with tropopauses at British overseas stations has shown that, unless rigorous definitions are adhered to, the values obtained in regions of double tropopause may vary from station to station in a quite spurious manner; it would be unwise therefore to regard the cross-sections as indicating more than the general tropopause level.

The supplementary cross-section for low latitudes at 165 E (Figs. 7, 11, 15, 19) used the following stations:-

(Na)	Nauru	00°32'S.	166°55'E.	R/W	April only
(Ku)	Kusaie	05°20'N.	163°05'E.	R/W	" "
(Po)	Ponape	06°50'N.	155°12'E.	P.B.	January only
(Kw)	Kwajalein	08°43'N.	167°44'E.	R/W.	
(En)	Eniwetok	11°21'N.	162°21'E.	R/W.	
(Wa)	Wake	19°18'N.	166°37'E.	R/W.	

The locations of the cross-sections which have been drawn have been dictated by the distribution of observing stations. It must be emphasized that they do not provide a representative selection of longitudes, since each coincides with a trough in the mean atmospheric circulation. It has not

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been possible here to construct cross-sections in the two main ridges of the circulation, the Atlantic and Mid-Pacific. Mid-Pacific cross-sections have, however, been drawn (3) (4) (9), and the cross-section for 0° longitude by Johnson (10) is probably characteristic of the Atlantic in winter. In such regions the movement of jet-streams is freer than elsewhere, and their speed less. Cross-sections may therefore be expected to show less pronounced maxima than are found in mean troughs. Also it seems probable that in general the area of the Westerlies will extend further southwards (see below).

3. CROSS-SECTIONS FOR INDIVIDUAL DAYS

The mean cross-sections for 80°W (Figs. 5, 9, 17) have drawn attention to a wind maximum found south of the polar jet-stream. Its presence has been known for some time (11), but no attempt to describe its structure seems to have been made.

To provide information about this, cross-sections for particular days of January and April (Figs. 21, 22) were drawn for 80°W. In the main, these used the same stations as the mean cross-sections. Where necessary, geostrophic winds derived from the potential temperature isopleths were used to supplement observed winds.

Fig. 23 is a cross-section for a day in July for 80°W. and demonstrates the easterlies in the stratosphere.

4. 200 mb. WIND DIAGRAMS

The speed and direction of the 200 mb. winds on each day of a month, for a selection of stations, have been plotted to provide a quick and simple comparison of high level flow in various regions of the world (Figs. 24-27). This method of presentation has been chosen because of its simplicity and because it shows the persistence of the winds, a quality which may prove to be very important in tropical meteorology, especially to the forecaster.

DISCUSSION

1. THE SUBTROPICAL JET-STREAM

At all seasons of the year mean charts of winds or contours reveal the presence of a belt of strong westerly winds, which has become known as the subtropical jet-stream, around the Northern hemisphere. It is at about 30°N. in winter and further North in summer. Though observations from the southern hemisphere are not nearly so complete, a similar belt is almost certainly present there (3), (9), (12).

JANUARY

In the cross-section along 80°W. for January 1951 (Fig. 5), the subtropical jet-stream appears at about 25°N, to the south of the polar jet-stream maximum and the observations from Miami (Fig.24(a)), Havana and San Juan show that between 40,000 and 50,000 ft. wind speeds in this area may be considerably in excess of 100 knots. The height of the maximum, approximately 160 mb., is higher than is found at other longitudes; at 45°E. and at 140°E., for example, it occurs at about 190 mb.

It is known that, in the mean, the polar jet-stream turns North-eastwards from America and is found on the South side of the Icelandic low (10). Though no upper air ascents are available to determine the course of the sub-tropical jet-stream directly, it appears probable that it also is continuous, at least on some days. Thus Dakar (Fig.24(b)), though well to the South of the anticipated centre, sometimes has strong winds at 200 mb.

Considerable strengthening of the stream occurs in its passage across North Africa and Arabia, so that at Bahrein the mean wind speed at 200 mb. has reached 100 knots. The 200 mb. winds at Bahrein and Miami (Figs.24(a) and (c)) show how much more variable in strength is the wind in the Caribbean area than

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in the Middle East. This is partly due to the greater variability of the positions of the wind maxima at 80°W, but it is also a reflection of the greater breadth of the Middle East stream, and of the fact that the flow over America is usually in two parts with lighter winds between.

Little information is available at high levels over the main Asian landmass in 1951, but cross-sections for other years have been published. Chaudhury (13) dealing with January and February at 76°E, shows the sub-tropical jet-stream at 31°N, just south of the Himalayas. He indicates a second maximum north of 40°N, and another at 15°-20°N whose presence does not seem to have been substantiated. In a winter cross-section for 120°E., Yeh (14) places the centre of the stream at 28°N, and shows a second maximum at 42°N. He comments upon the very small day-to-day variation of the position of the maximum wind. Bolin (15) has suggested that the double maximum found by both Chaudhury and Yeh is produced by the splitting of the subtropical stream by the Himalayas.

The stream emerges from the Asian landmass just North of 30°N (Figs. 1 and 8) with a southerly component and a mean speed at the centre of nearly 200 knots. In another investigation (5), where the month of January 1951 was also considered, another maximum is shown at 40°N. Though the presence of a comparatively ill-defined low level maximum on the 400 mb. surface is confirmed by the present work, the author does not regard the evidence for a very pronounced maximum at 200 mb. as conclusive.

Over the Pacific the jet-stream can be detected in the observations for Hilo (Fig.24(d)) and Mokuleia (21°35'N 158°10'W). The character of the stream, however, probably changes very considerably in this region; the centre is weaker than over Japan, and its position considerably more variable. Also, there appears to be a suggestion in the observations that the stream is split, (Fig. 1) one centre being near Hawaii, the other about 10-15° further North.

APRIL

In April (Figs. 2, 9-12), the subtropical jet stream is further North and its speed less than in January. Its height on all three cross-sections is 180 mb. approx. On the whole, however, the differences between the two months are quite small.

Over Central America, the stream is at 28°N (Fig. 9). The wind maximum produced by the polar jet-stream is, for this month, rather ill defined but probably is present just to the North of 40°N. This is purely a characteristic of 1951; for in some years e.g. 1949, 1950, the position of the maximum may be readily fixed from the mean 300 mb. contours.

As in January, the sub-tropical stream can be traced across Africa, Asia (Fig.25(b)), Japan (Fig.25(c)), and the Pacific (Fig.25(d)). Over the Pacific its behaviour is again far from clear but the suggestion that there are two centres, one just South of Hawaii (20°N), the other between 25-30°N, is once more present (Fig. 2). The observations made at Mokuleia and Hilo (Fig. 25(d)) are remarkable because of the very pronounced mean horizontal shear between them, and the marked mean vertical shear at the former station. The following is a summary of the 300 and 200 mb. winds from the two stations for April 1951:-

/MOKULEIA

MOKULEIA (21°58'N 159°22'W)

200 mb. APRIL, 1951.

Speed (kt.)	DIRECTION 0°	30°	60°	90°	120°	150°	180°	210°	240°	270°	300°	330°
0-9	.	.	.	.	.	.	.	.	1	.	.	.
10-19	.	.	.	.	.	.	.	.	.	1	1	.
20-29	.	.	.	.	.	.	.	.	.	1	.	.
30-39	.	.	.	.	1	.	.	.	.	2	2	.
40-49	1	.	.	.	.	.	.	.	.	2	1	1
50-59	.	.	.	.	.	.	.	.	1	.	1	.
60-69	.	.	.	.	.	.	.	.	.	1	1	.
70-79	.	.	.	.	.	.	.	.	1	1	.	.
80-89	.	.	.	.	.	.	.	.	.	2	.	.

No. of observations - 22
Mean Wind - 43 knots/270°.

300 mb. APRIL, 1951.

	0°	30°	60°	90°	120°	150°	180°	210°	240°	270°	300°	330°
0-9	.	1	.	.	.	.	.	.	.	.	1	.
10-19	1	1	1	.	.	.	1	.	.	2	.	.
20-29	2	.	1	1	.	.	.	.	2	1	2	.
30-39	.	.	.	.	.	.	.	.	.	2	.	.
40-49	.	.	.	1	.	.	.	.	.	.	1	.
50-59	.	.	.	.	1	.	.	.	.	2	.	.

No. of observations - 24
Mean Wind - 9 knots/290°.

HILO (19°44'N 155°04'W)

200 mb. APRIL, 1951.

	0°	30°	60°	90°	120°	150°	180°	210°	240°	270°	300°	330°
0-9	.	.	.	.	.	.	.	.	.	1	.	.
10-19	.	.	.	.	.	.	.	.	.	1	.	.
20-29	.	.	.	.	.	.	.	.	.	.	1	.
30-39	.	.	.	.	.	.	.	.	.	1	.	.
40-49	.	.	.	.	.	.	.	.	.	.	2	1
50-59	.	.	.	.	.	.	.	.	2	3	2	.
60-69	.	.	.	.	.	.	.	.	.	3	1	.
70-79	.	.	.	.	.	.	.	.	.	1	.	.
80-89	.	.	.	.	.	.	.	.	1	2	.	.
90-99	.	.	.	.	.	.	.	.	.	3	.	.
100-109	.	.	.	.	.	.	.	.	.	1	.	.

No. of observations - 26
Mean Wind - 58 knots/270°.

300 mb. APRIL, 1951.

	0°	30°	60°	90°	120°	150°	180°	210°	240°	270°	300°	330°
0-9	.	.	.	.	.	.	.	.	.	1	.	.
10-19	2	.	.	.	.	.	.	.	2	.	1	.
20-29	2	.	.	.	.	.	.	1	.	3	1	.
30-39	.	.	.	.	.	.	.	.	1	4	1	.
40-49	.	.	.	.	.	.	.	1	.	1	.	.
50-59	.	.	.	.	.	.	.	.	.	4	2	.
60-69	.	.	.	.	.	.	.	.	.	.	.	.
70-79	.	.	.	.	.	.	.	.	.	1	.	.

No. of observations - 28
Mean Wind - 30 knots/270°.

/While

While it is possible that there are errors in the reported winds for Mokuleia (extracted from the Daily Series Synoptic Weather Maps published by the U.S. Weather Bureau) there is no apparent specific reason for doubting their accuracy, and the value must be accepted as indicating that the winds in the Pacific show characteristics which are peculiar to that region, and which have not yet been explained.

JULY

In July, the position of the subtropical jet-stream is generally between 35°and 40°N, and it is considerably weaker than in either January or April (Figs.3, 13-16). Its height is also less than for the earlier months. Thus for 140°E the maximum is at 200 mb. and in the Middle East the observations from Nicosia, which was probably very near the centre of the stream, show that the maximum was probably below this level.

Over America its position, estimated from the mean 300 mb. contours published by the U.S. Weather Bureau, is about 45°N i.e. somewhat further North than for other longitudes. This is apparently owing to the formation in July of a region of high pressure (evident on the 300 and 200 mb. mean contours) over the Mexico desert, causing the stream to be diverted Northwards.

At longitude 45°E the maximum is apparently at about 40°N, while at 140°E it is just south of this latitude. Over the Pacific, its course is uncertain; it is difficult to establish continuity between the winds over Japan and the Westerlies at Johnson (16° 44'N, 169° 31'W) and Hawaii, and it may be that the stream is broken in Mid-Pacific (Fig. 3).

OCTOBER

The data available for October are not as complete as for the other months. They show the subtropical jet-stream returning southwards towards its winter position, and increasing in strength (Figs. 4, 17-20). (In general its position and speed are intermediate between the April and July values). The height at which it occurs seems to vary with longitude; at 80°W. the centre is at about 180 mb., at 45°E. it is at about 170 mb. and at 140°E. it is at about 200 mb.

At 80°W. (Fig. 17) the stream is comparatively weak, though the observations from Havana (Fig. 27(a)) show that quite high wind speeds may occur. The maximum here is well to the south of the polar jet-stream, which, according to the mean 300 mb. contours, had its centre at about 50°N.

In the Middle East (Figs. 18, 27(b)) the speed of the stream is again strong and fairly constant, while over Japan (Figs. 20, 27(a)), where it reaches its highest speed, the winds are apparently more variable than in the Spring or Winter. In the Pacific its position is apparently North of Hawaii, but no observations are available to determine it.

STRUCTURE OF THE SUBTROPICAL JET-STREAM

Figs. 21 and 22 have been drawn for individual days of January and April. The main conclusions which may be drawn from them about the structure of the subtropical jet-stream in the Caribbean region are:-

(1) The stream is not associated with a frontal system; on most occasions examined it was centred over a surface high pressure system.

(2) The stream occurs at a considerably greater height than jet-streams of mid-latitudes, usually about or somewhat above 200 mb. Further, there is generally no evidence of its presence on the 500 mb. wind field, and it may not be apparent even at 300 mb.

(3) Though the stream occurs entirely within the tropical air mass, the atmospheric thermal structure which produces it seems to be associated with the presence of a lower tropopause to the North. There seems to be a tendency for

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the main vertical wind shear to be found immediately above a special point on the R/S ascent which is approximately isentropic with the lower tropopause and where the isentropic surfaces are steeply sloping. This requires further explanation. To maintain the strong baroclinity as a stable feature, there must, presumably, be a meridional movement of air, and it seems possible that the region is one in which a tropospheric-stratospheric interchange of air takes place.

2. THE EASTERLIES

Equatorwards of the Westerlies there is a region of Easterly winds, whose extent varies with the N-S movement of the subtropical jet-stream.

JANUARY

Considering only the Northern hemisphere, the area of Easterlies is a minimum in January; at the same time, the subtropical jet-stream reaches its maximum velocity and most Southerly position (Fig. 1). The most marked feature of the variation of the area of Easterlies with longitude is the rapid decrease which occurs in the Western Pacific; thus a comparison of the cross-sections for 140°E and 165°E shows the Westerlies farther South at 165°E (Figs. 7, 8). Still farther East, at Canton (a pilot balloon station 02°49'S 171°43'W), winds have a Westerly component between 300 and 150 mb. (Fig. 1) and it seems that in this region there may in the mean be no Easterly winds at high levels.

There is a marked change in the extent of the Easterlies with height. It is greatest near the surface and decreases to a minimum at about 150-200 mb. (Figs. 5-8). The cross-sections show that the Westerly maximum occurs at about the same level. Though the area of the Easterlies increases above 150 mb., Westerlies predominate in the stratosphere.

A typical ascent through the Easterlies would show a maximum below about 450 mb., then decreasing Easterlies to 200-150 mb. (probably becoming Westerly at these levels) followed by increasing Easterlies up to about the tropopause. This form of wind variation with height is found at almost all stations in January. Exceptions are Nairobi and Singapore which give a high level easterly maximum. At the former, this occurs near 150 mb., and at the latter, near 100 mb. High level maxima are also found at these stations in July, when conditions near the Equator may be expected to be similar to those in January.

Though the way in which these winds fit into the general flow will probably remain in doubt until more observations are available from the Southern hemisphere, there seems to be evidence here for the existence at some periods of the year of a high altitude equatorial Easterly stream. It is true that in July the high level maxima could be considered as being part of the strong Easterly stream produced across Africa and the South of Asia by the heating of the landmass to the North (see below), and a similar stream across Southern Africa could be responsible for the Nairobi January maximum. But this would leave unexplained the January maximum at Singapore, which has no large landmass to the South.

Even if an equatorial stream does exist, it is clear that it is not continuous around the Equator, for Canton (02°49'S, 171°43'W) winds do not show a high level tropospheric maximum.

APRIL

The Easterlies extend farther North than in January (Fig. 2), but in most respects the two months are very similar. It is notable, however, that Nairobi now shows the more usual variation of wind with height with a maximum at 450 mb., while at Singapore the mean wind increases slowly to 300 mb. and is then almost constant to 100 mb.

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In the Western Pacific, the area of Easterlies is again smaller than elsewhere, and a feature, the stability of which it is not possible to judge from one year's data, is that at most levels the Westerlies increase southward South of Wake for a limited distance (Fig. 11). This is also found in July (Fig. 15 and see below). The 200 mb. winds at Kwajalein (Fig. 25(f)) are quite constant in direction, and it does not therefore seem possible to explain the increase as being produced by a few weather systems giving strong Westerlies in the vicinity of Kwajalein.

The winds at Kwajalein may be contrasted with those for Nauru (00°32'S 166°55'E) where during March and April of 1951, wind measurements were made rather irregularly. The following is a summary of the 200 mb. observations for the two months:-

	Direction (° from North)											
	0	30	60	90	120	150	180	210	240	270	300	330
SPEED (kt.)												
0-9	.	.	.	.	1	.	.	.	.	.	.	.
10-19	.	1	2	1	1	.	.	.	.	.	1	.
20-29	.	.	1	.	1	2	2	.	1	.	.	.
30-39	1	.	.	.	.	.	.	.	.	.	.	.
40-49	.	.	.	.	1	1	.	.	.	.	.	.
50-59	.	.	.	.	.	1	1	.	.	.	.	.

No. of observations - 19
Mean wind - 15 knots/135°.

The stronger winds were scattered through the period; they were not produced by one particular weather system. The variation in speed and direction is most marked. With such a distribution, it is doubtful whether much real significance can be attached to the value of the mean wind. It may be expected to change considerably from year to year. No observations are available from Nauru for other seasons, so it cannot be stated whether such conditions are characteristic of the months March-April, or whether they prevail throughout the year. It seems probable, however, having regard to the observations made elsewhere near the Equator, that January and July show a smaller variation.

JULY

The area of Easterlies is much more extensive in July than in the preceding months of the year (Fig. 3). Many stations now have an Easterly maximum at high levels.

A strong steady Easterly stream is found at Aden (16) (Fig.26(b)) where the maximum occurs at about 175 mb. The thermal field suggests that the centre of the stream is North of Aden (Fig. 14) and that the maximum velocity there occurs at about 120 mb., (although it is not legitimate to use the geostrophic relation quantitatively at this latitude, the thermal field probably still gives the correct sense of the variation of wind with height). The stream can be traced back across India, to the south of Hong Kong, thence over the Pacific Ocean where it appears as a long arm of Easterlies stretching into the general Westerly flow (Fig. 3). Westwards from Aden the stream can be detected at Lagos. At most of the stations affected by it the wind increases through the tropopause and presumably continues to increase until the maximum of the stratospheric Easterlies is reached (see below). Bahrein, Hong Kong (17), Clark Field and Marcus Island all show this form of variation of wind with height. At 60 mb., the Easterlies at Bahrein are comparable in strength with those at Aden. Lagos, however, has a tropospheric maximum and it seems probable that thereafter the wind decreases to give Westerlies or only light Easterlies in the stratosphere (as for Nairobi Fig.(14), and see below).

There is no evidence for a strong Easterly stream across Central America (Fig. 26(a)), though Havana and Miami show Easterlies which increase through

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the tropopause and they may be produced by a similar but very much weaker stream; nor can such Easterlies be detected over the oceans, other than the Western Pacific. It is concluded that the Easterly stream is caused by the heating up of the Asian landmass in summer.

Fig. 23 shows the stratospheric Easterlies. Though constructed for a particular day, it is a good approximation to the mean, since the time variation of winds in the stratosphere during July is small. Stratospheric Easterlies are found in almost all latitudes in summer at a height which increases with latitude. Over the Equator in July, Westerlies are found at high levels (Fig. 14) and the centre of these Easterlies must therefore occur some way North of the Equator.

As already mentioned under "January", Nairobi and Singapore show a maximum in the troposphere. At Nairobi it is at about 200 mb. and weaker than in January; at Singapore it is again at 100 mb., and stronger than in January.

Easterlies are strong at 200 and 150 mb. at both Hong Kong and Singapore (Fig. 3) but are noticeably weaker at Clark Field (15°10'N 120°34'E). It has already been suggested that there are two Easterly centres, one South of Hong Kong, the other at or near the Equator. To the East of Asia both streams weaken, and at 165°E (Fig. 15), the wind between them has become Westerly, giving a Westerly centre embedded in an Easterly flow. Still farther East only a narrow belt of equatorial Easterlies remains.

The Caribbean cross-section for July (Fig. 13) shows Westerlies over San Juan, which admit a similar explanation. The heating over the Arizona desert in Summer produces a warm centre which at high levels has an anticyclonic circulation around it, and this gives Easterly winds at Miami and Havana. South of Havana, Westerlies are again found before the true equatorial Easterlies commence. Presumably over the Atlantic, where there is no continental influence, Westerlies can be expected to be maintained at 200 mb. at least as far South as 17°N.

It has been suggested (18), (19), (20) that the Westerlies embedded in Easterlies are produced by cyclonic or anticyclonic systems which move in a general Easterly current, so as to give Westerlies consistently along the same track. This explanation is not necessarily in conflict with that given above, since it may be the manner in which the influence of the Asian and American landmasses are made evident. It should be pointed out, however, that Hubert(21) has discussed Westerlies which occurred at high altitudes in the Western Pacific in September, and which had no compensatory Easterlies elsewhere; in view of the similarities found between the atmospheric circulation in Summer and Autumn, it seems quite possible that such Westerlies could be present in July.

OCTOBER

In October, the Easterlies appear, in the mean, to be featureless (Figs. 17, 19), though it is not known how much this may be due to the smaller number of data available for this month. The area of Easterlies in the Northern hemisphere is less than in July, but is greater than in January or April.

Easterlies which increase with height in the upper troposphere and lower stratosphere are again present though the speeds are much smaller than in July. Clark Field has such winds, and the Easterly stream across the South of Asia is probably present in a weak form. Singapore still has a maximum at 100 mb. though it is considerably less pronounced than in either January or July; on the other hand Nairobi has Westerlies, which are stronger than those found in April, at 200 mb.

The figures for Wake and Eniwetok (Figs.27(f), 27(e)) suggest that the low speed of the resultant mean winds, and the general absence of systematic variation of the mean values within the Easterlies, may be caused by the predominance of moving vortices at high levels, producing Easterly and Westerly winds in almost equal numbers. It may be significant in this connection that in the Pacific,

/October

October is one of the months favourable for typhoon development (22).

3. YEAR-TO-YEAR VARIATION

Enough information about high level winds has not yet accumulated to enable one to see the problem of the year-to-year variation of the mean flow in proper perspective. Many stations have been making observations for only a short time, and 1950 is probably the first year for which sufficient observations were made to allow the pattern of the flow at low latitudes to emerge.

However, it is already clear that, in some areas at least and contrary to early expectations a considerable variation exists. Thus, considering the July observations for the Pacific, we find that, though Wake in 1950 (Fig.26(f)) gave almost entirely Westerly winds at 200 mb. and a mean wind of over 20 knots, in 1951 (Fig. 26(e)), Easterlies are found on about half the days and the resultant was a calm. Other stations in the area also indicate that in 1951, the Easterlies in the Pacific were stronger than in the previous year.

CONCLUSION

The observations of high level winds now available in the Northern Hemisphere are sufficient to indicate the mean flow over a considerable area of the tropics and subtropics. For the southern hemisphere, observations becoming available from Australia and South Africa will soon provide valuable information of the mean flow there.

It is clear that this investigation should be considered as only a first attempt to summarise the mean flow at high levels in tropical and subtropical regions. As more observations become available and as other years are considered, modification of the picture presented will almost certainly be necessary.

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APPENDIX.

The following is a list of the stations used in Figs. 1-4 :-

	Position	Months
0-20°E.		
Maison-Blanche	36°43'N. 03°14'E.	Jan.
Malta	35°50'N. 14°27'E.	Jan. Apr. Jly. Oct.
¤ Lagos	06°35'N. 03°20'E.	" " " "
20°-40°E.		
Nicosia	35°09'N. 33°17'E.	" " " "
Benina	32°05'N. 20°16'E.	" " " "
≠ Nairobi	01°17'S. 36°50'E.	" " " "
40°-60°E.		
Habbaniya	33°22'N. 43°34'E.	" " " "
Bahrein	26°16'N. 50°37'E.	" " " "
Aden	12°50'N. 45°01'E.	" " " "
60°-80°E.		
New Delhi	28°35'N. 77°12'E.	Apr. Jly. Oct.
Karachi	24°53'N. 67°07'E.	Apr.
Udaipur	24°35'N. 73°42'E.	Jly.
Bhuj	23°15'N. 69°48'E.	Jan.
Verewal	20°54'N. 70°22'E.	Apr.
Gadag	15°25'N. 75°38'E.	Jan.
Vengurla	15°15'N. 73°40'E.	Jan. Apr.
Bangalore	12°38'N. 77°35'E.	Jan.
Nagercoil	08°11'N. 77°25'E.	Jan. Apr. Jly. Oct.
80°-100°E.		
Dibrujah	27°28'N. 94°55'E.	Oct.
Raipur	21°14'N. 81°39'E.	Jan.
Masulipatam	16°11'N. 81°08'E.	Jan. Jly. Oct.
Madras	13°04'N. 80°15'E.	Apr.
Port Blair	11°40'N. 92°43'E.	Apr. Oct.
100°-120°E.		
Hong Kong	22°18'N. 114°10'E.	Jan. Apr. Jly. Oct.
≠ Singapore	01°23'N. 103°42'E.	" " " "
120°-140°E.		
Tateno	36°03'N. 140°08'E.	Apr. Jly. Oct.
Yonago	35°26'N. 133°21'E.	Jan.
Kagoshima	31°34'N. 130°33'E.	Jan. Apr. Jly. Oct.
Okinawa	26°12'N. 127°45'E.	Jly.
Clark Field	15°10'N. 120°34'E.	" " " "
Yap	09°24'N. 138°08'E.	Jan. Apr.

/140°

¤ The Lagos observations are for 1952. The following table compares the mean wind components at Nairobi for January, July and October of 1951 and 1952, from which the relevance of the 1952 Lagos observations may be judged. Too few observations are available at Nairobi in April to make a comparison in that month.

	January				July				October				
	1951		1952		1951		1952		1951		1952		
	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	(knots)
300 mb.	-2	20	1	15	1	13	-2	1	3	3	-2	0	
200 "	-9	26	-9	18	4	22	3	4	-2	-6	-2	-11	
150 "	-10	34	-2	11	-3	8	2	16	-4	-7	-3	-9	
100 "	1	17	-2	4	1	8	0	26	1	5	-2	4	

≠ The April mean wind at Nairobi is for 1952.

≠ The Singapore mean winds are for the period 1950-52(1).

	Position		Months			
140°-160°E.						
Iwo Jima	24°47'N.	141°20'E.	Jan.	Apr.		
Marcus Island	24°17'N.	153°58'E.		Apr.	Jly.	Oct.
Guam	13°27'N.	144°44'E.	"	"	"	"
(including NORTHFIELD and HARMON FIELD).						
Ponape	06°50'N.	155°12'E.	Jan.	Apr.		
160°-180°E.						
Wake	19°18'N.	166°37'E.	Jan.	Apr.	Jly.	Oct.
Eniwetok	11°21'N.	162°21'E.	"	"	"	"
Kwajalein	08°43'N.	167°44'E.	"	"	"	"
Majuro	07°06'N.	171°24'E.		Apr.		
Kusaie	05°20'N.	163°05'E.		Apr.		
Nauru	00°32'S.	166°55'E.		"		
Nandi	17°45'S.	177°27'E.	Jan.	Apr.	Jly.	Oct.
Whenuapai	36°47'S.	173°38'E.	"	"	"	"
Ohakea	40°12'S.	175°23'E.	Jan.		Jly.	
180°-160°W.						
Midway	28°13'N.	177°21'W.		Apr.	Jly.	
Johnson Is.	16°44'N.	169°31'W.	Jan.	Apr.	Jly.	Oct.
Palmyra	05°52'N.	162°06'W.		Apr.		
Canton	02°49'S.	171°43'W.	Jan.	Apr.	Jly.	Oct.
160°-140°W.						
O.W.S. 24	28°N.	145°W.	"	"	"	"
Mokuleia	21°35'N.	158°10'W.	Jan.	Apr.	Jly.	
Hilo	19°44'N.	155°04'W.	"	"	"	"
140°-120°W.						
Oakland	37°44'N.	122°12'W.	Jan.		Jly.	Oct.
Santa Maria	34°56'N.	120°25'W.	Jan.	Apr.	Jly.	Oct.
O.W.S. 34	33°N.	135°W.	"	"	"	"
120°-100°W.						
Grand Junction	39°06'N.	108°32'W.	"	"	"	"
Albuquerque	35°03'N.	106°37'W.	Jan.		Jly.	Oct.
100°-80°W.						
Oklahoma City	35°24'N.	97°36'W.	Jan.	Apr.	Jly.	Oct.
Charleston	32°54'N.	80°02'W.	"	"	"	"
Burrwood	28°59'N.	89°22'W.	"	"	"	"
Brownsville	25°55'N.	97°28'W.	"	"	"	"
Miami	25°49'N.	80°17'W.	"	"	"	"
Havana	23°09'N.	82°21'W.	"	"	"	"
80°-60°W.						
Greensboro	36°05'N.	79°57'W.		Apr.	Jly.	
Hatteras	35°15'N.	75°40'W.	Jan.		Jly.	Oct.
Kindley Field	32°22'N.	64°40'W.	Jan.	Apr.		
San Juan	18°28'N.	66°06'W.	Jan.	Apr.	Jly.	Oct.
Albrook Field	08°59'N.	79°34'W.	"	"	"	"
20°W- 0°						
Gibraltar	36°09'N.	05°21'W.	"	"	"	"
// Dakar	14°40'N.	17°26'W.	Jan.			Oct.

// The October mean wind at Dakar is for 1950.

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FIGURE 1
MEAN WINDS FOR JANUARY

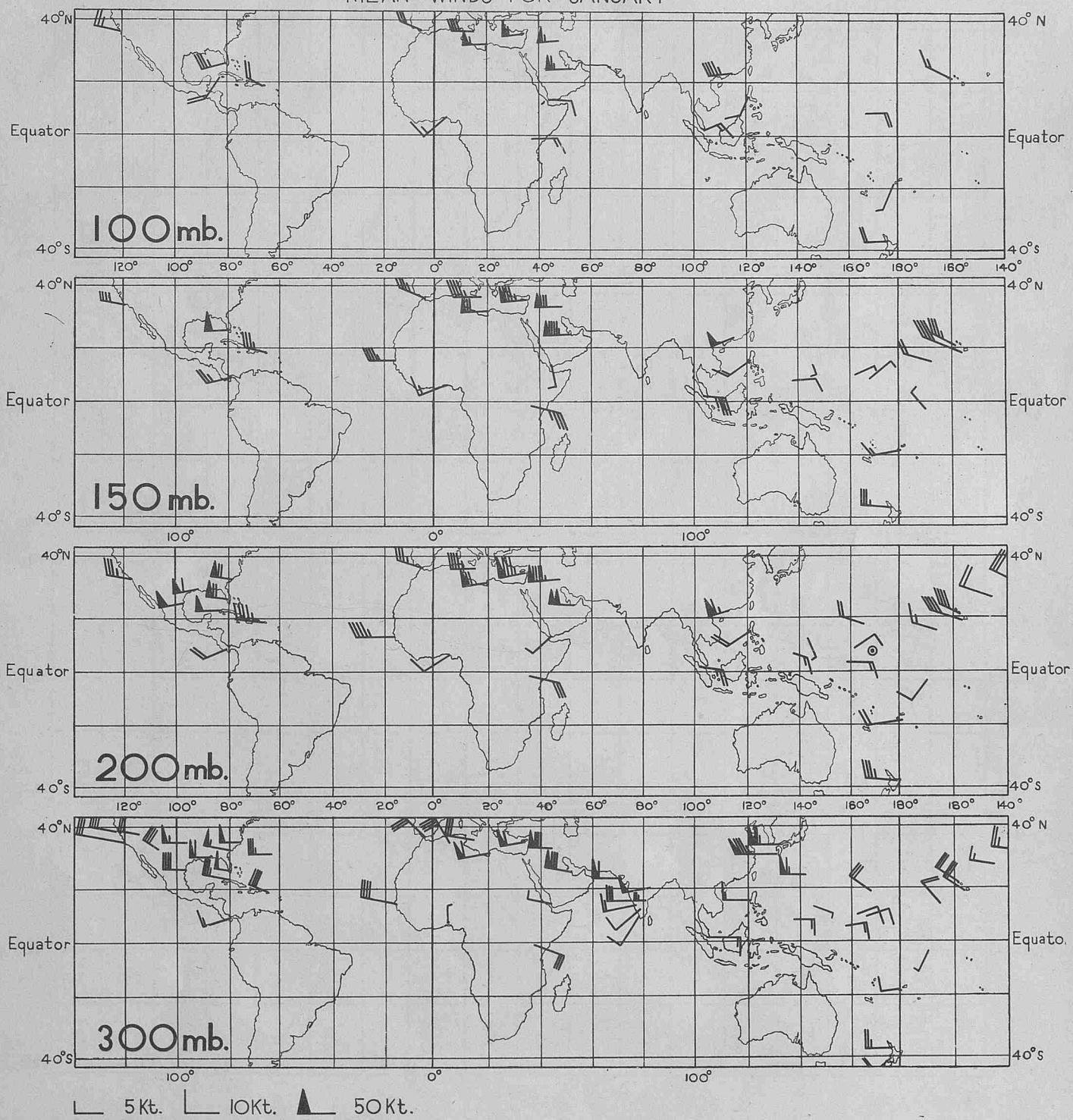

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FIGURE 2
MEAN WINDS FOR APRIL

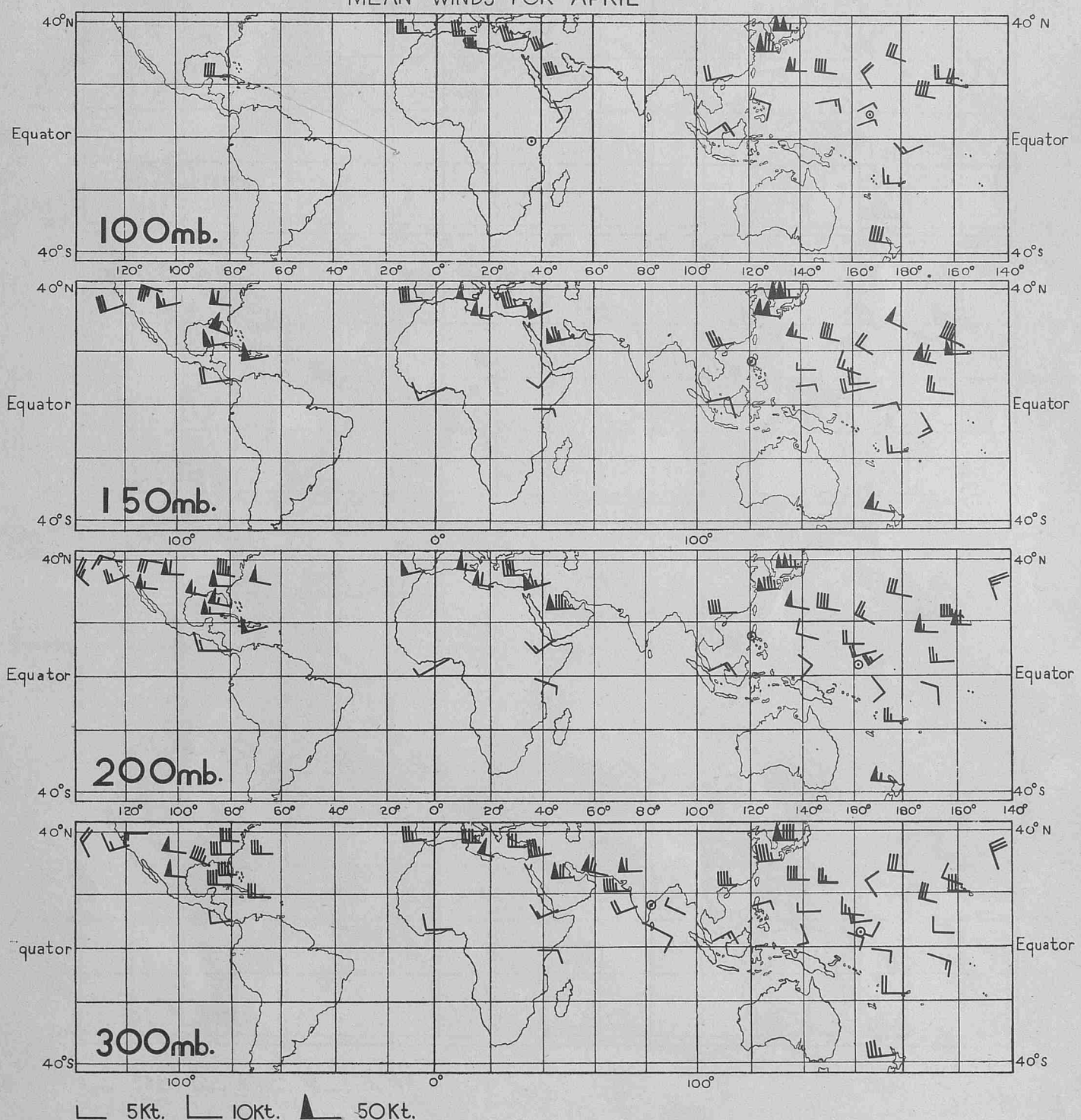

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FIGURE 3
MEAN WINDS FOR JULY

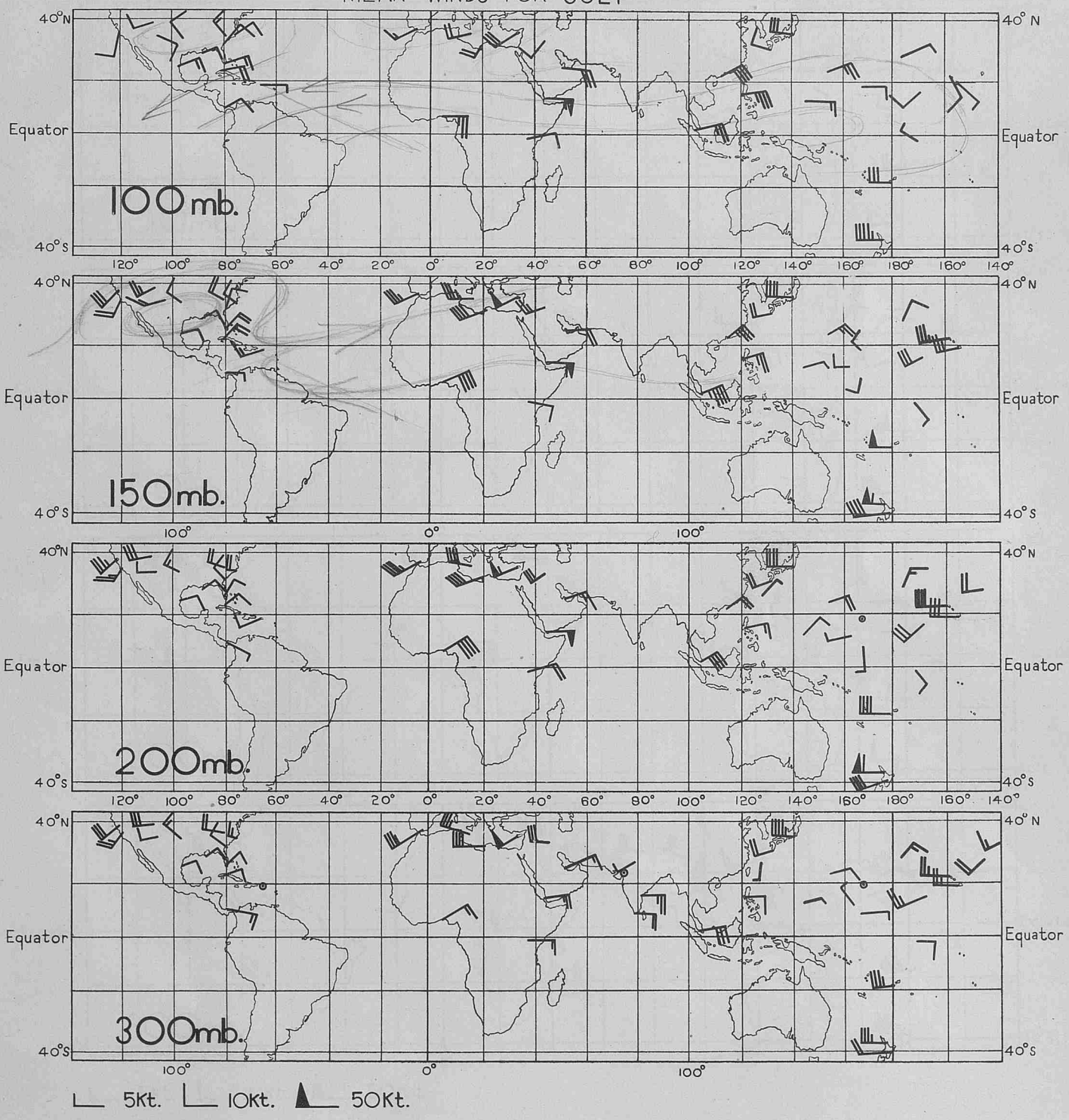

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FIGURE 4
MEAN WINDS FOR OCTOBER

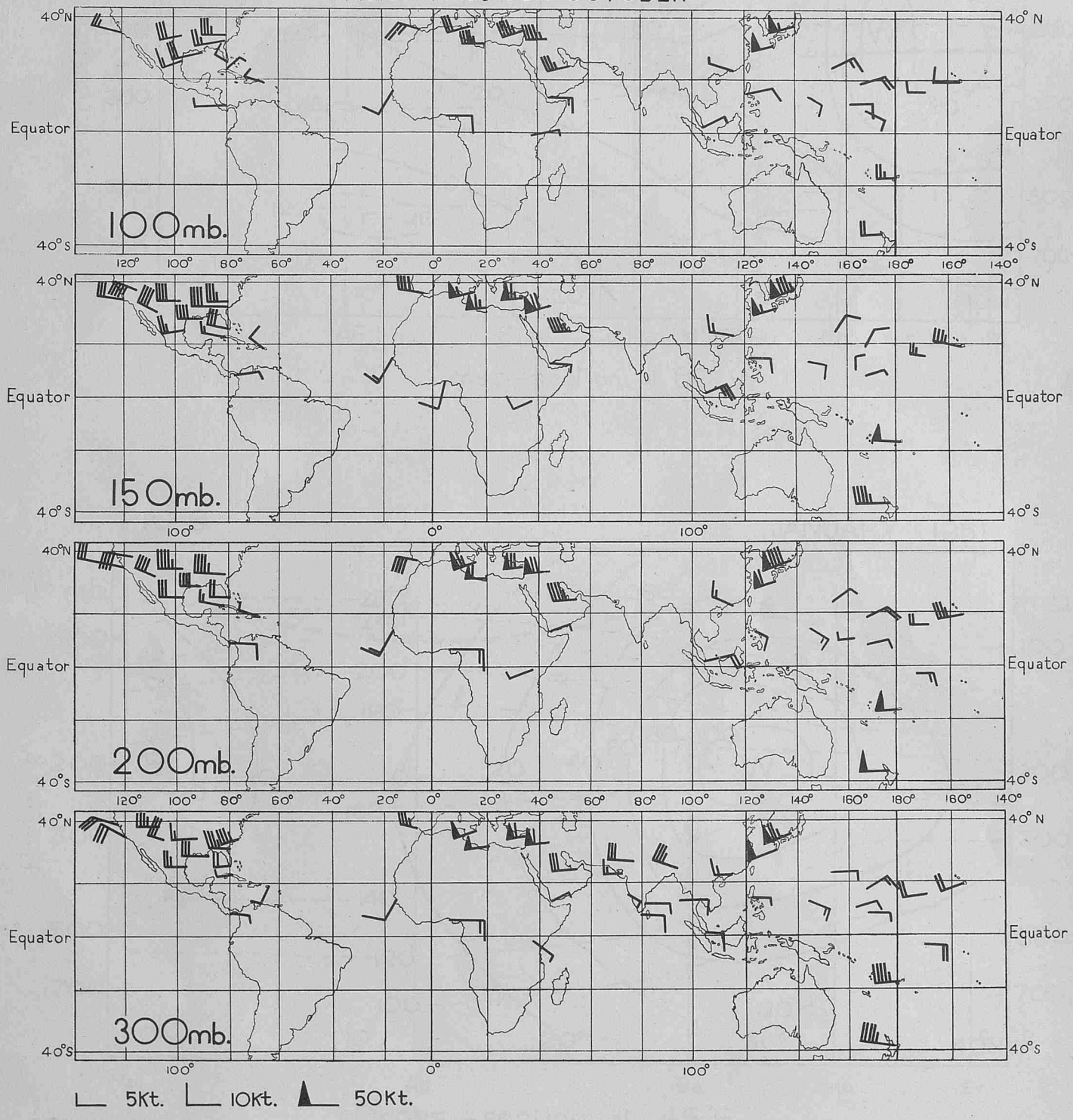
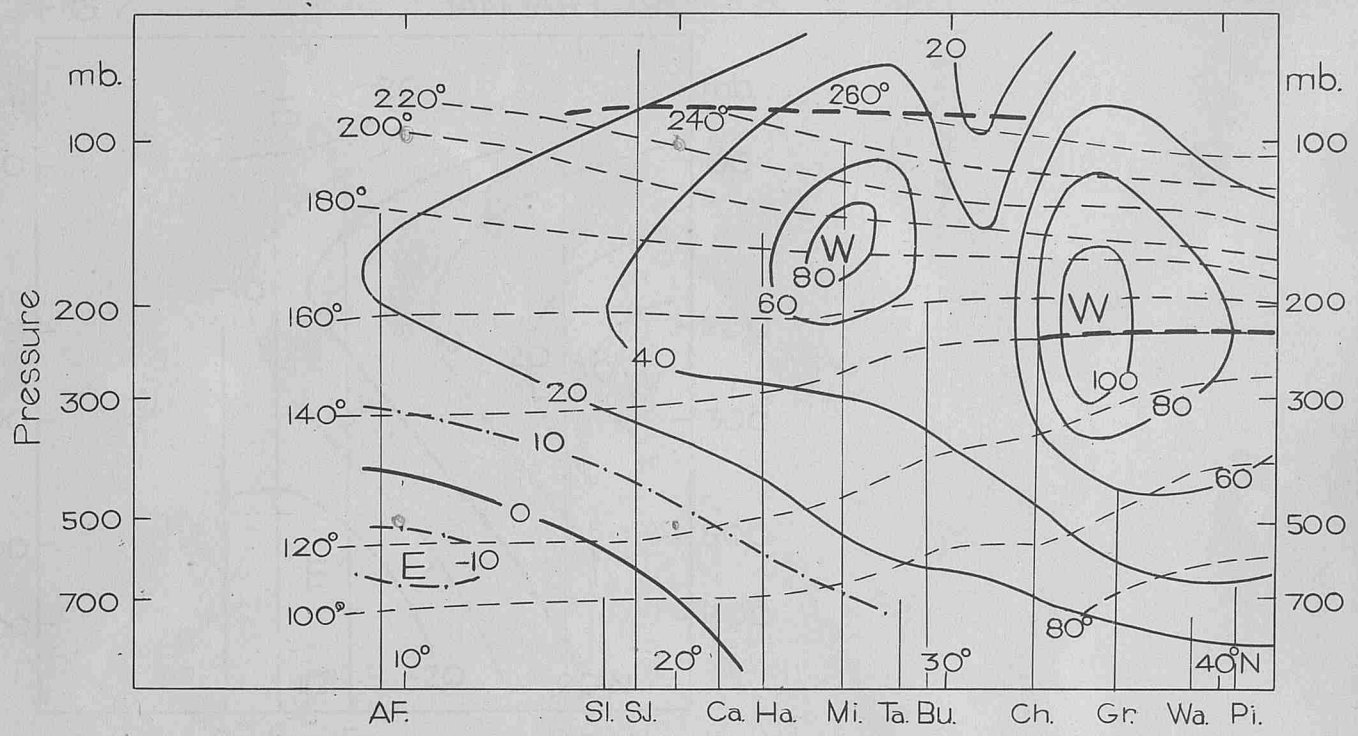
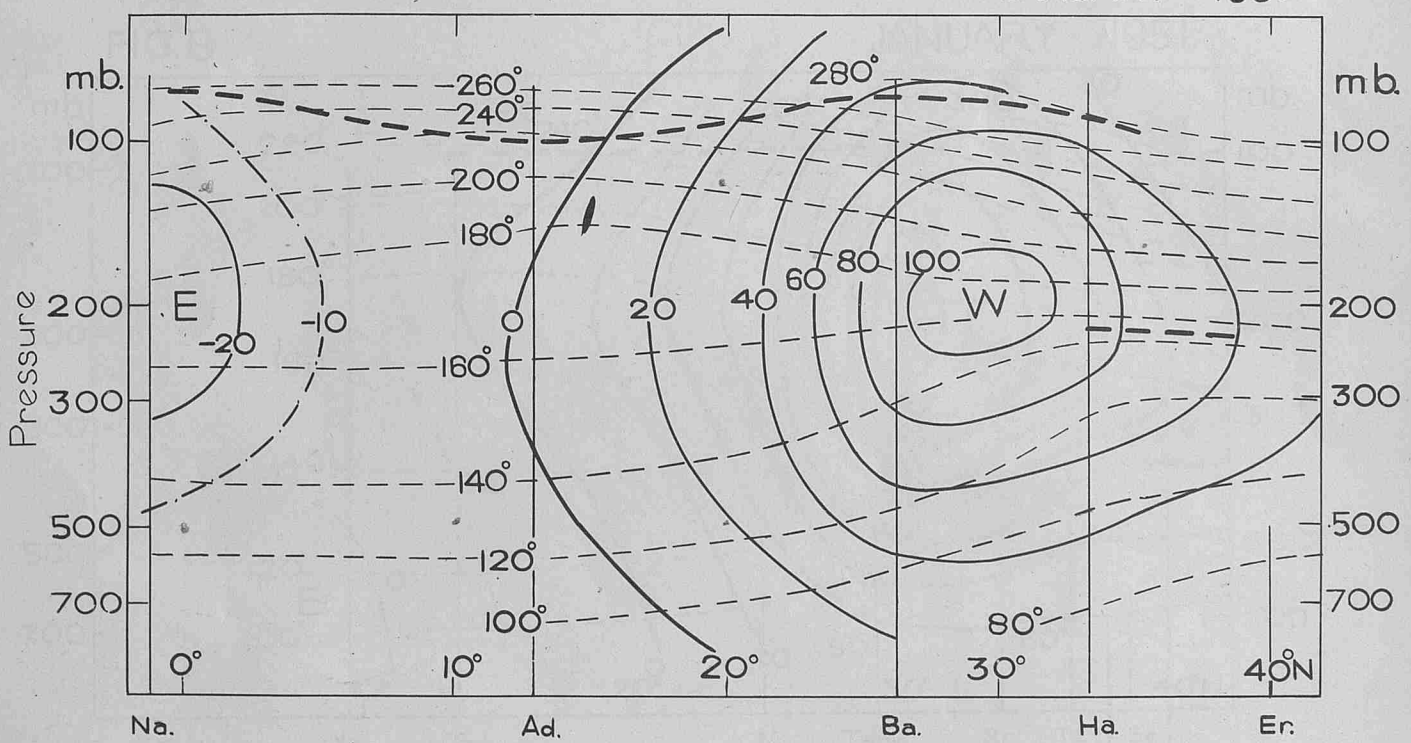


FIG.5 JANUARY 1951



Cross-section at 80° W

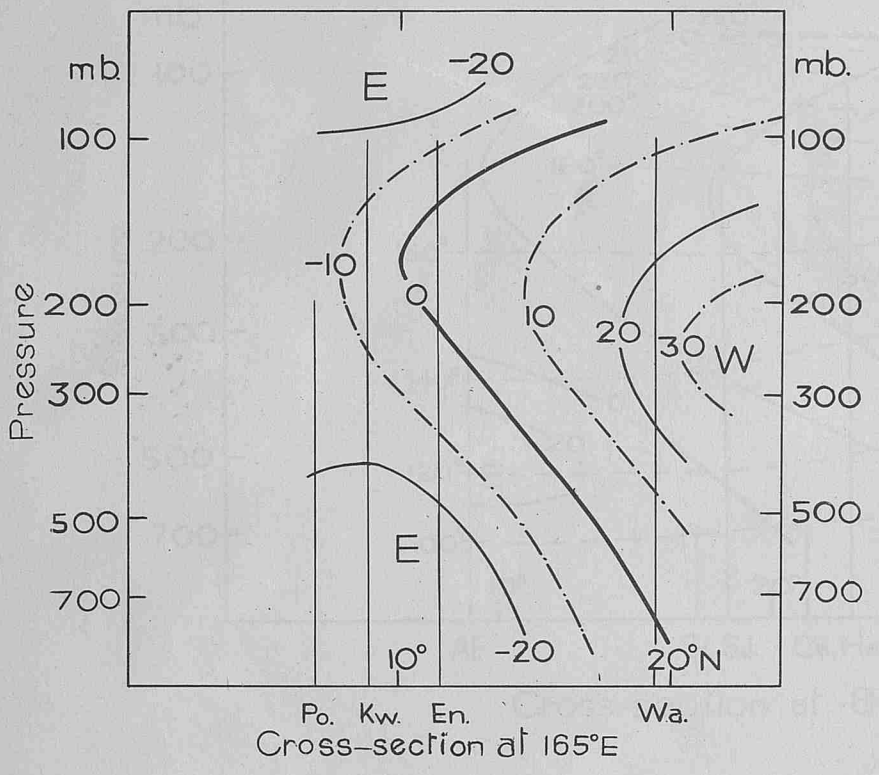
FIG. 6 JANUARY 1951



Cross—section at 45°E

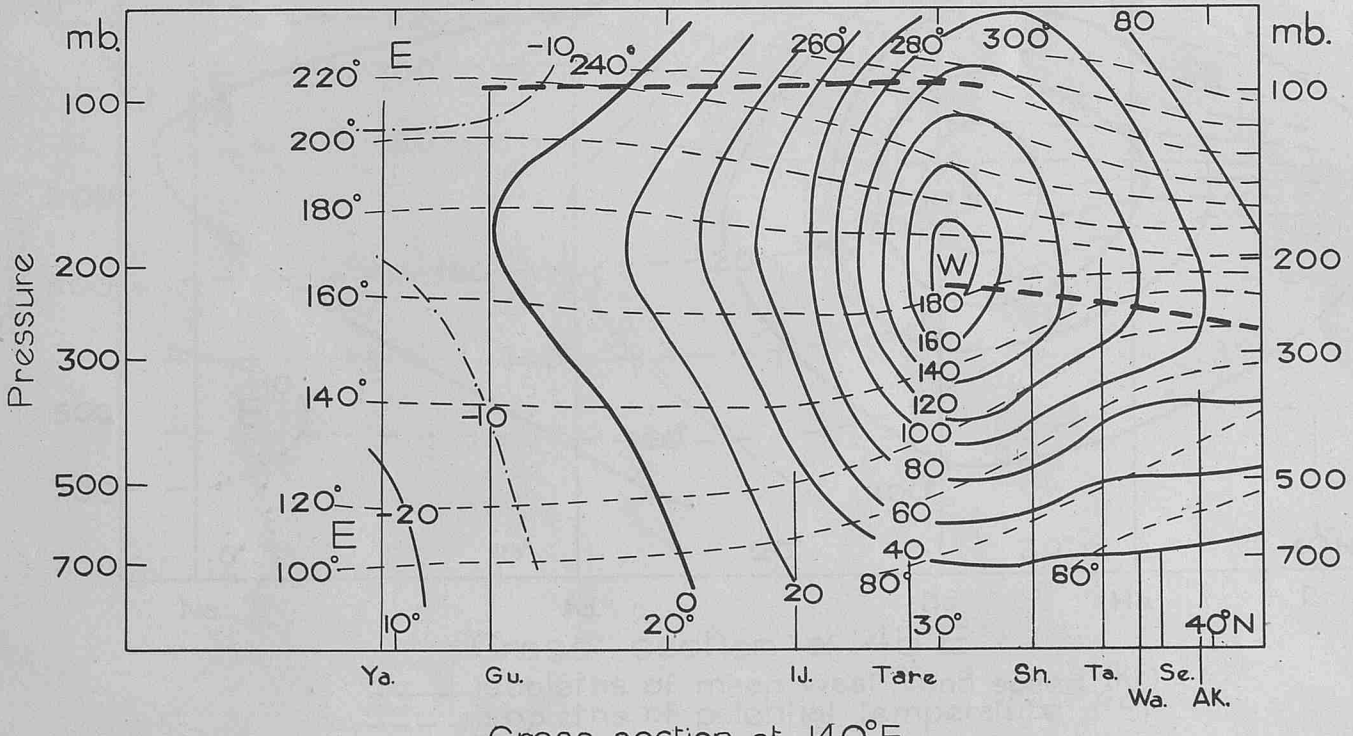
——— Isopleths of mean west wind speed (kt.)
– – – – Isopleths of potential temperature (°F.)
— — — Tropopause

FIG. 7 JANUARY 1951



Cross-section at 165°E

FIG. 8 JANUARY 1951



Cross-section at 140°E

——— Isopleths of mean west wind speed (kt)
– – – – Isopleths of potential temperature (°F.)
— — — Tropopause

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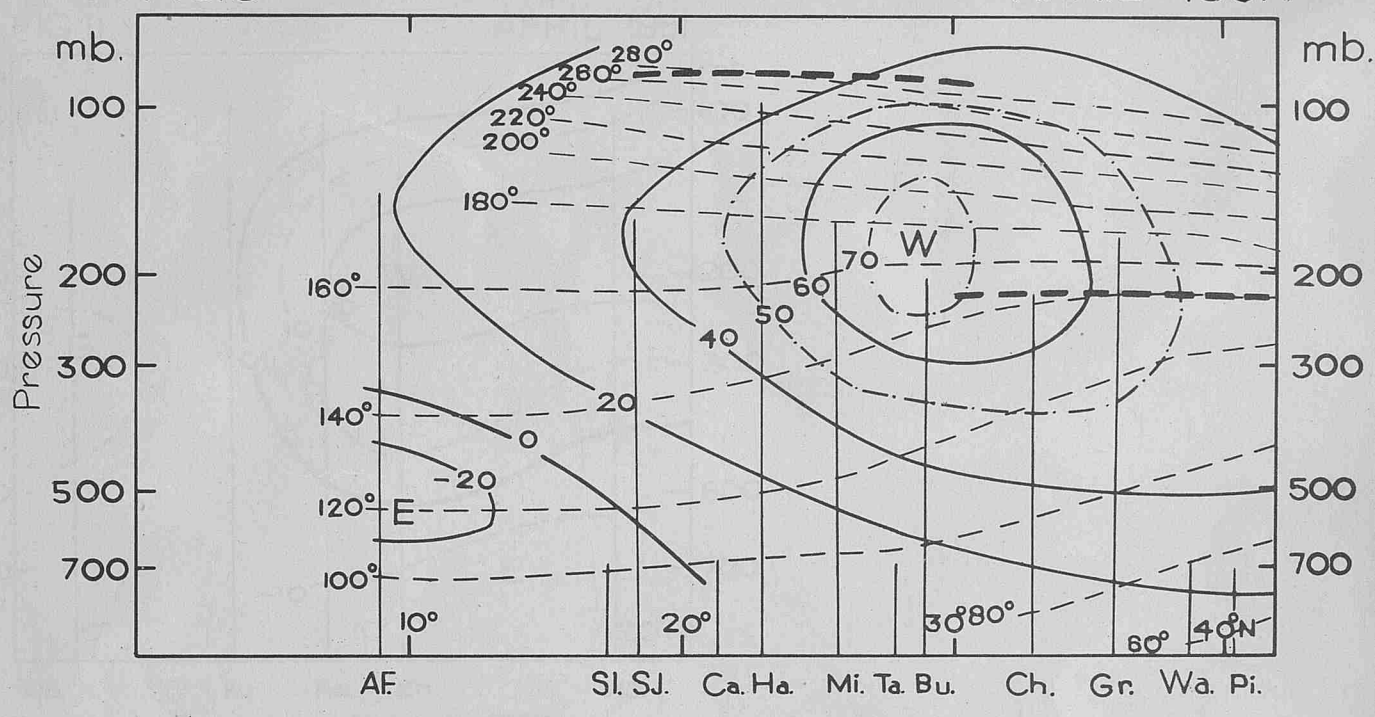
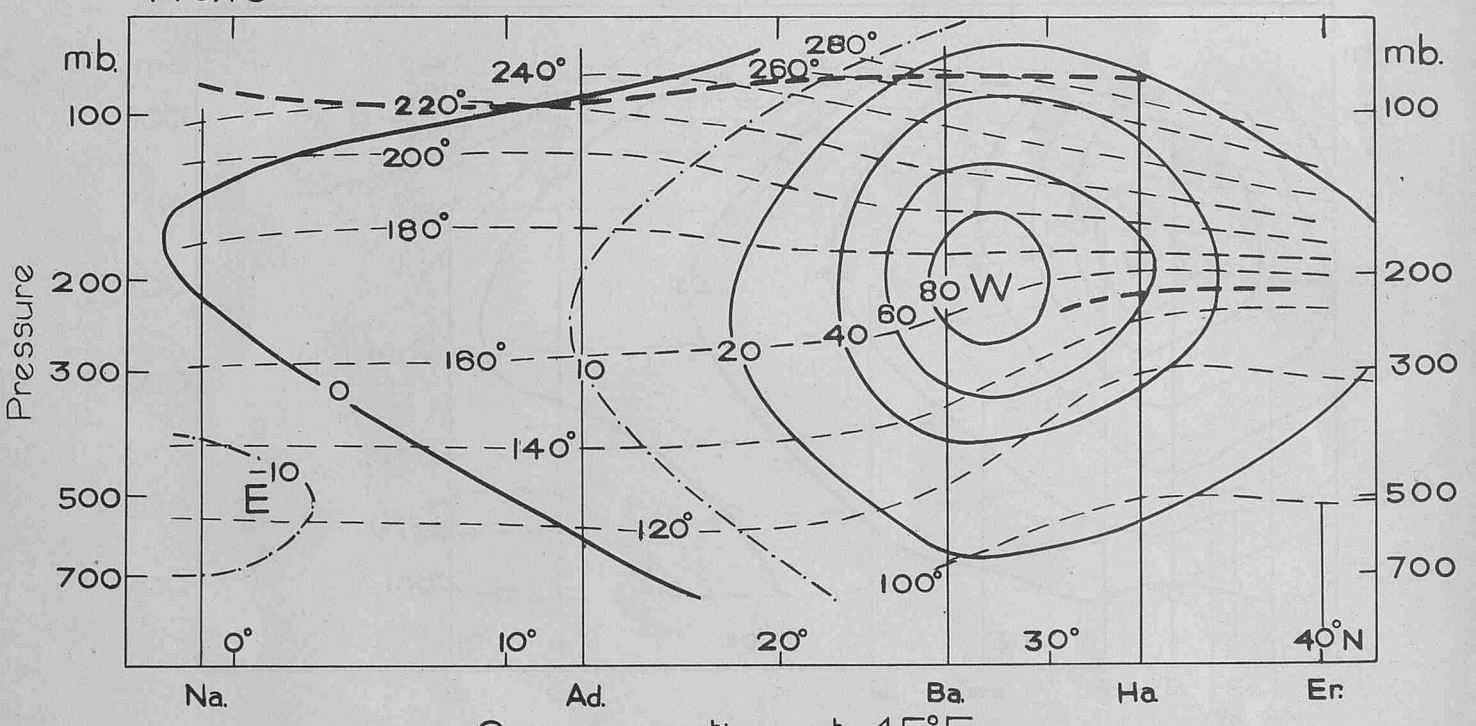
FIG. 9 APRIL 1951



Cross-section at 80°W

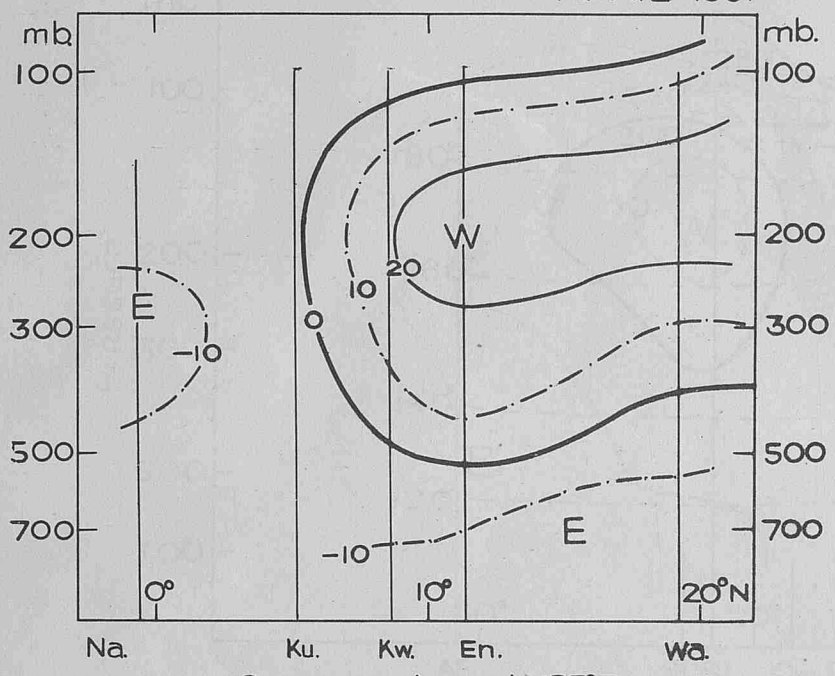
FIG. 10 APRIL 1951



Cross-section at 45°E

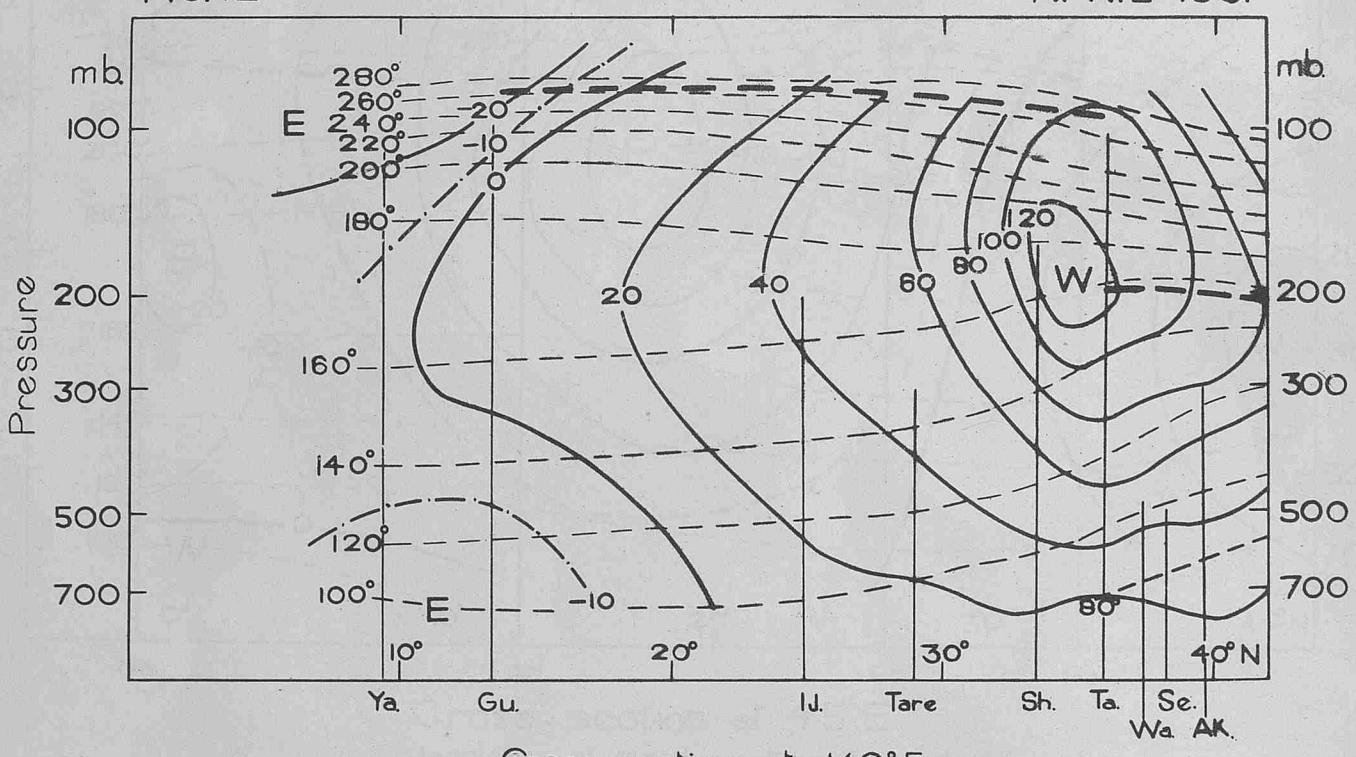
——— Isopleths of mean west wind speed (Kt.)
- - - - Isopleths of potential temperature (°F.)
— — — Tropopause

FIG. 11 APRIL 1951



Cross-section at 165°E

FIG. 12 APRIL 1951



Cross-section at 140° E

——— Isopleths of mean west wind speed (Kt.)

- - - - Isopleths of potential temperature (°F.)

— — — Tropopause

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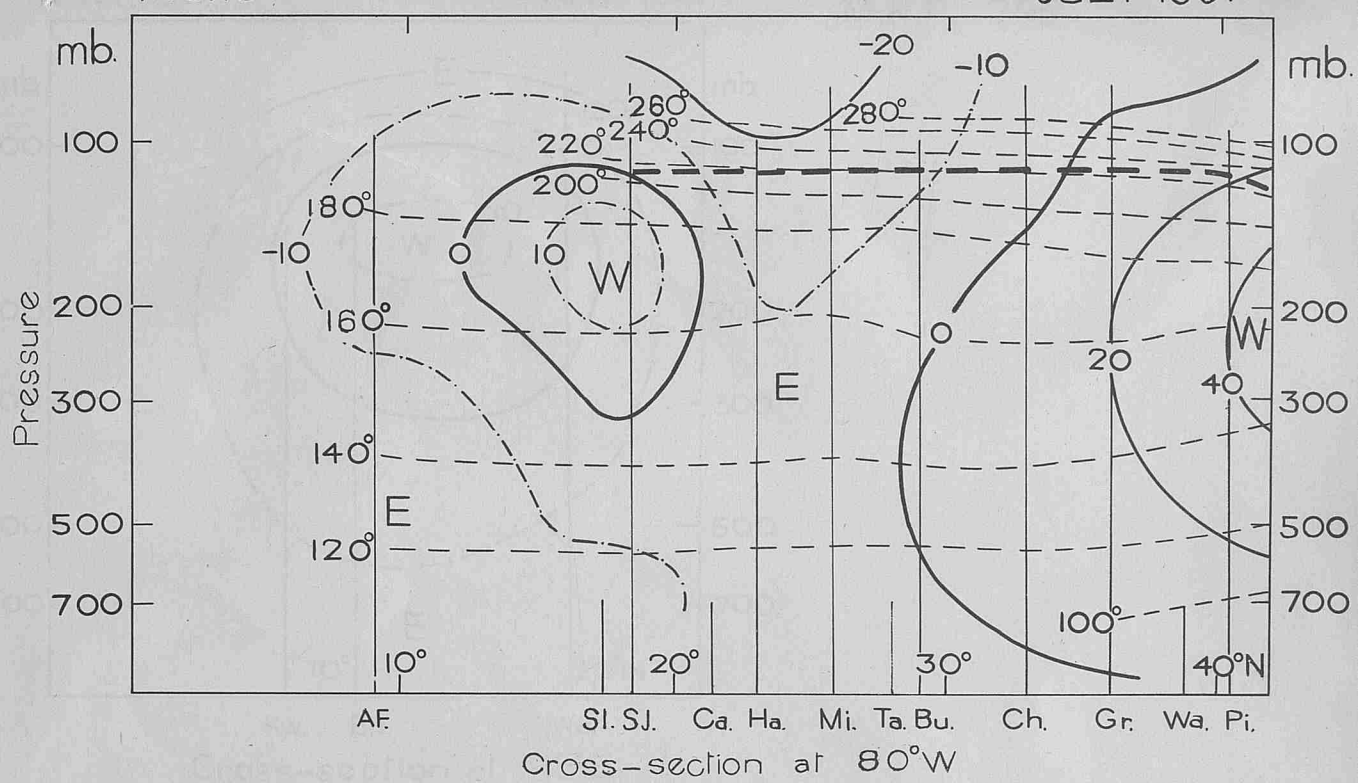
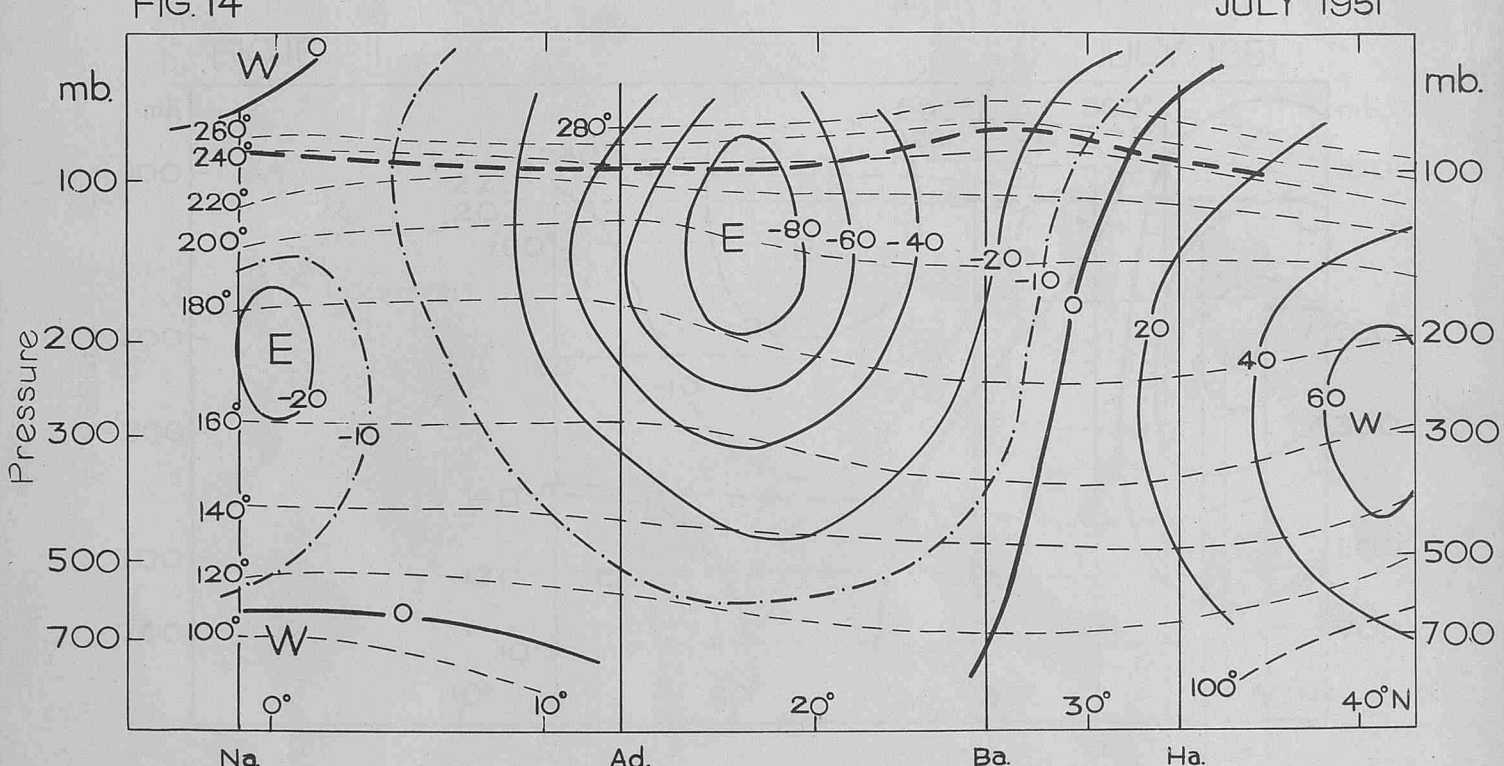
FIG. 13 JULY 1951



Cross-section at 80°W

FIG. 14 JULY 1951



Cross-section at 45°E

——— Isopleths of mean west wind speed (Kt.)

- - - - Isopleths of potential temperature (°F.)

— — — Tropopause

FIG. 15 JULY 1951

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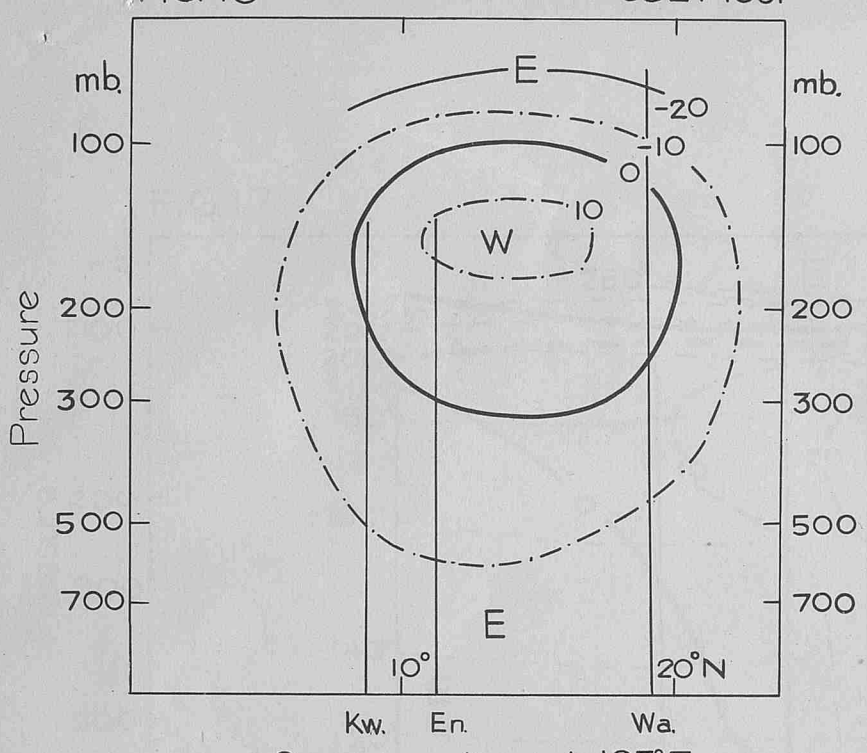
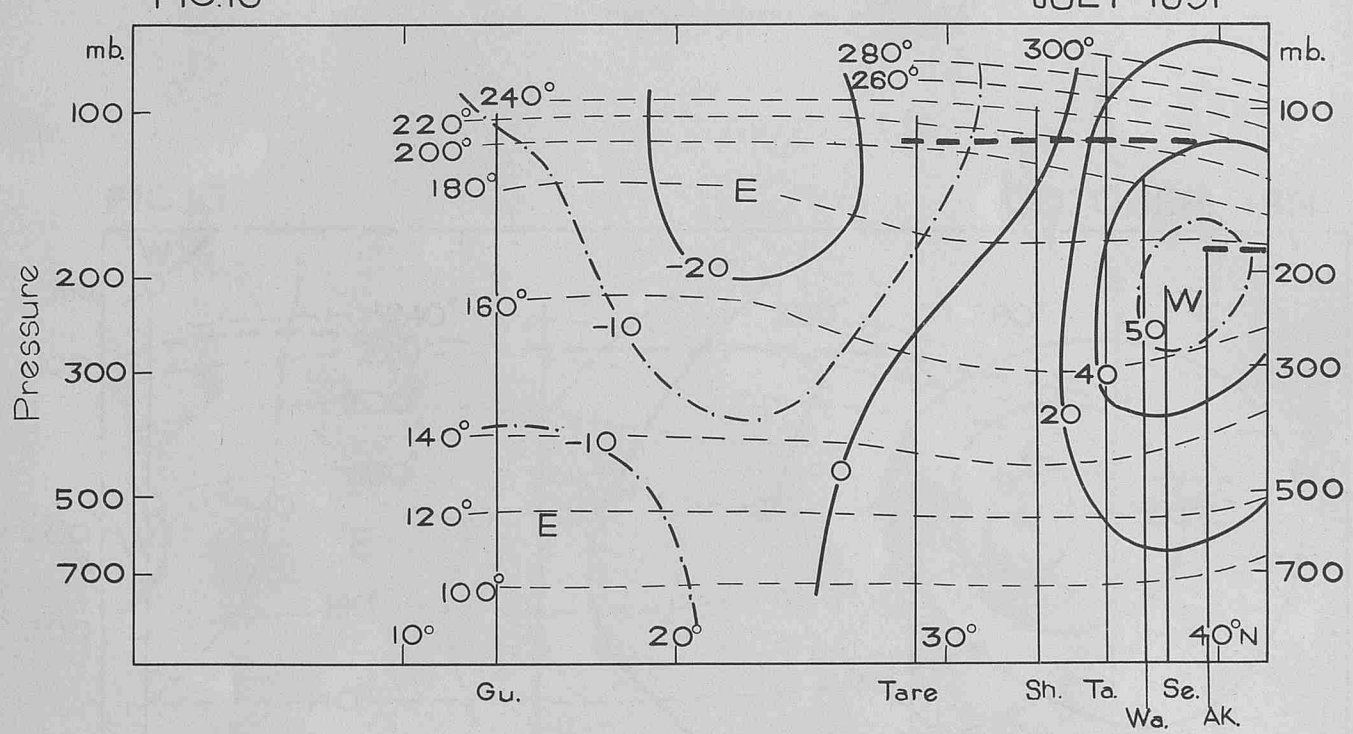


Cross-section at 165°E

FIG. 16 JULY 1951



Cross section at 140°E

——— Isopleths of mean west wind speed (K.t.)

– – – Isopleths of potential temperature (°F.)

— — Tropopause

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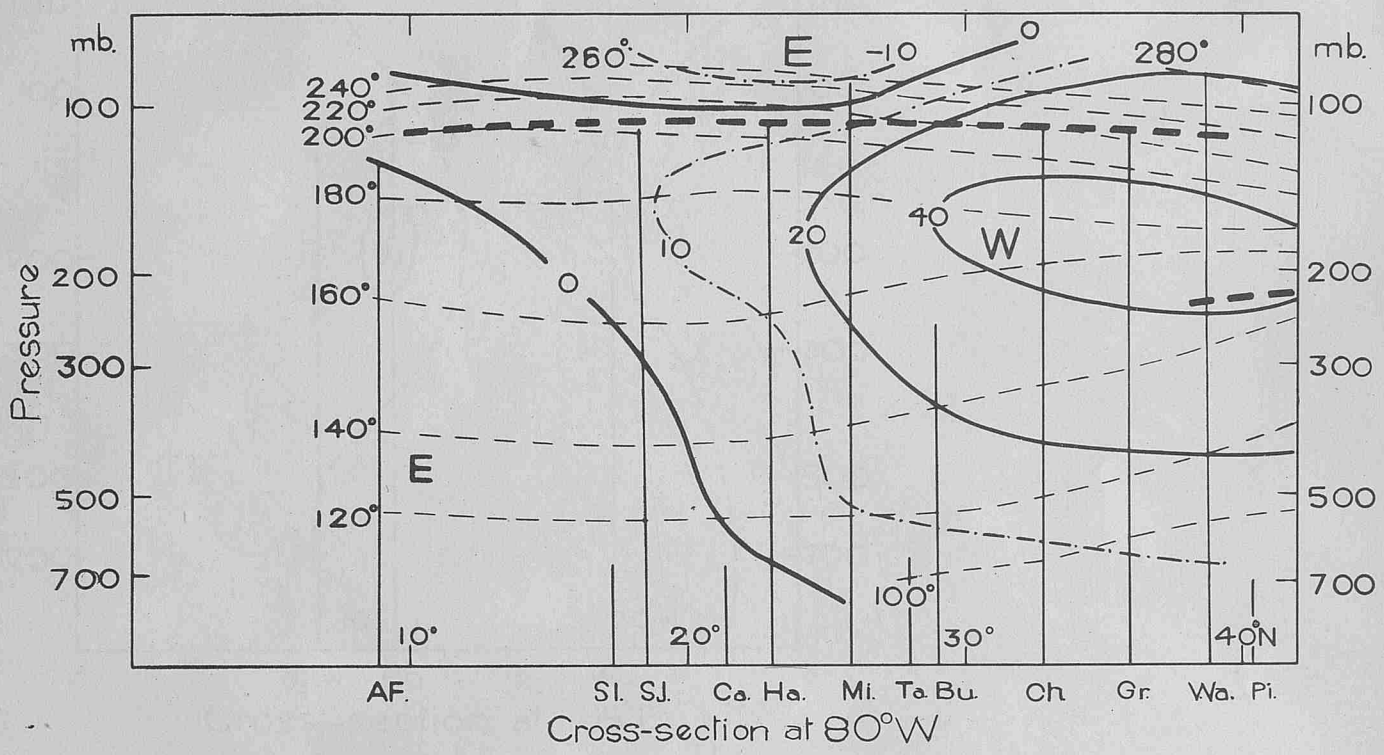
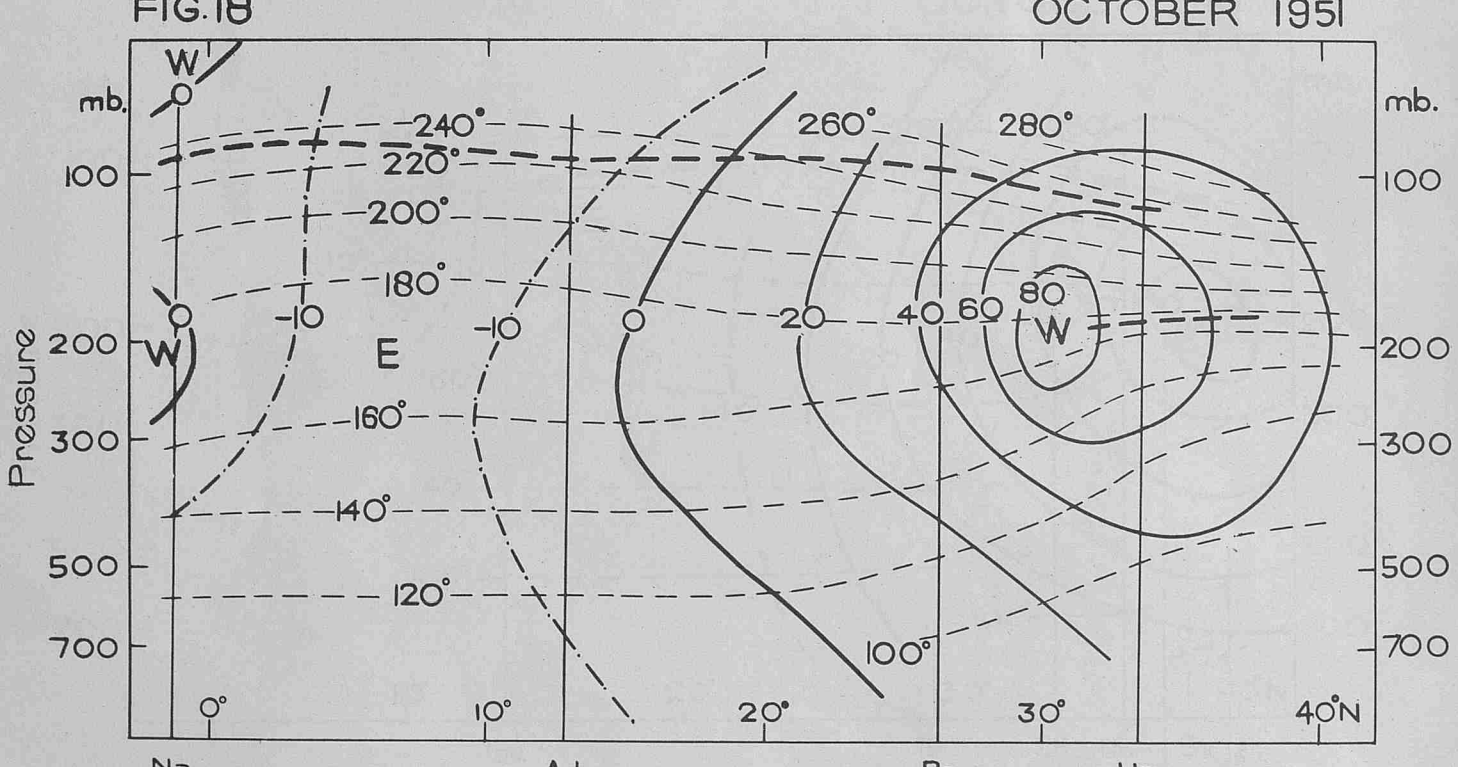
FIG. 17 OCTOBER 1951



Cross-section at 80°W

FIG. 18 OCTOBER 1951



Cross-section at 45°E

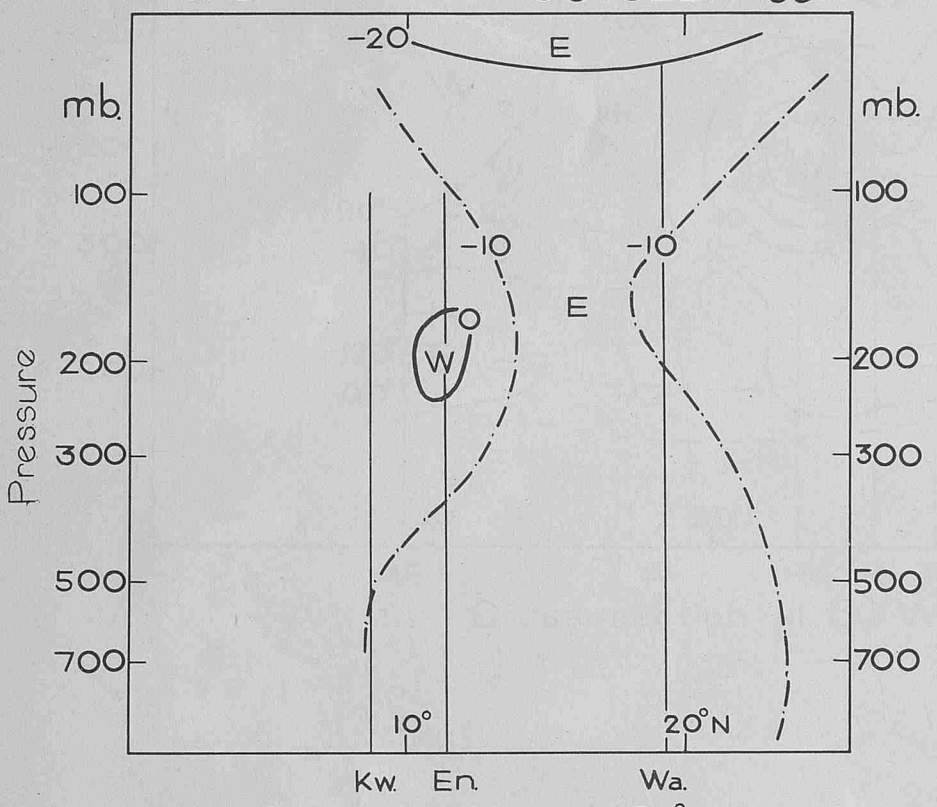
——— Isopleths of mean west wind speed (Kt.)

– – – Isopleths of potential temperature (°F.)

— — — Tropopause

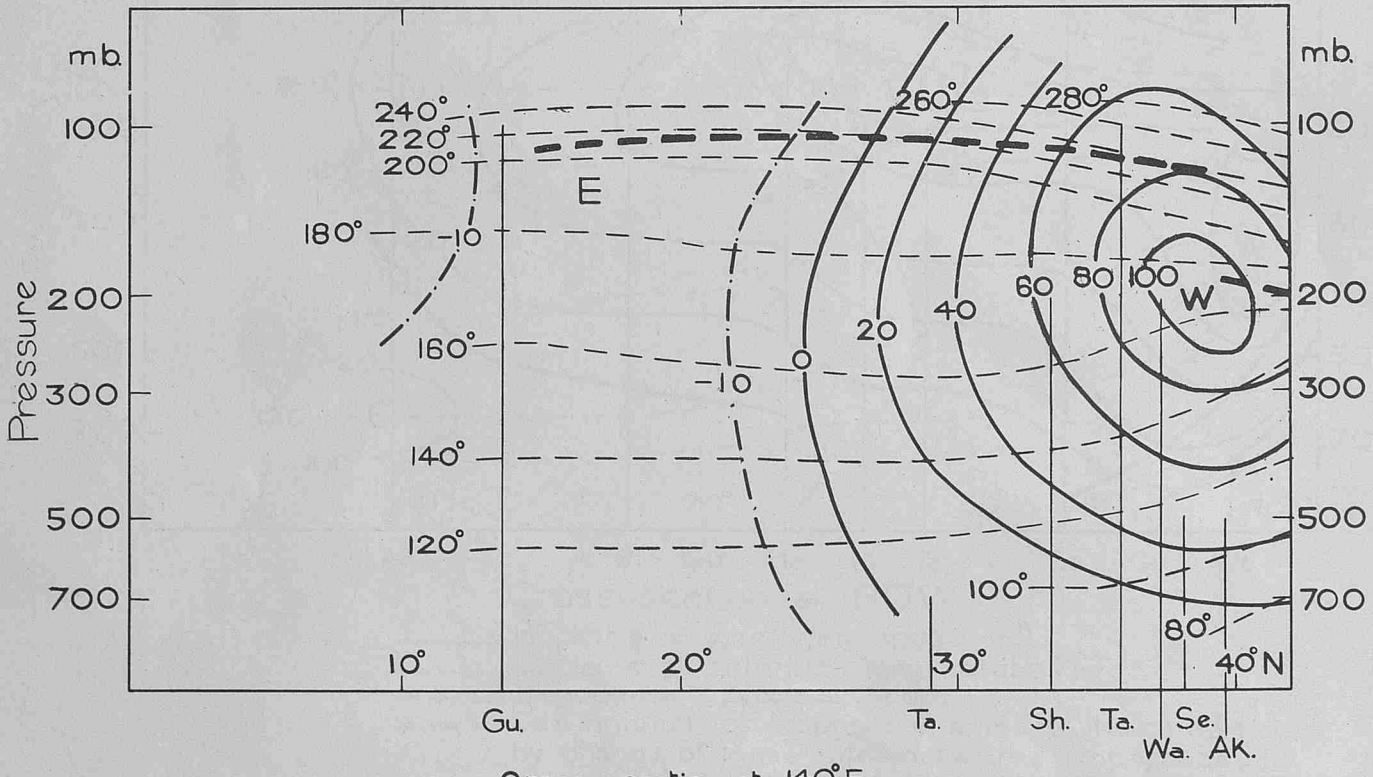
FIG. 19 OCTOBER 1951

M.R.P. 795



Cross-section at 165°E

FIG. 20 OCTOBER 1951



Cross-section at 140°E

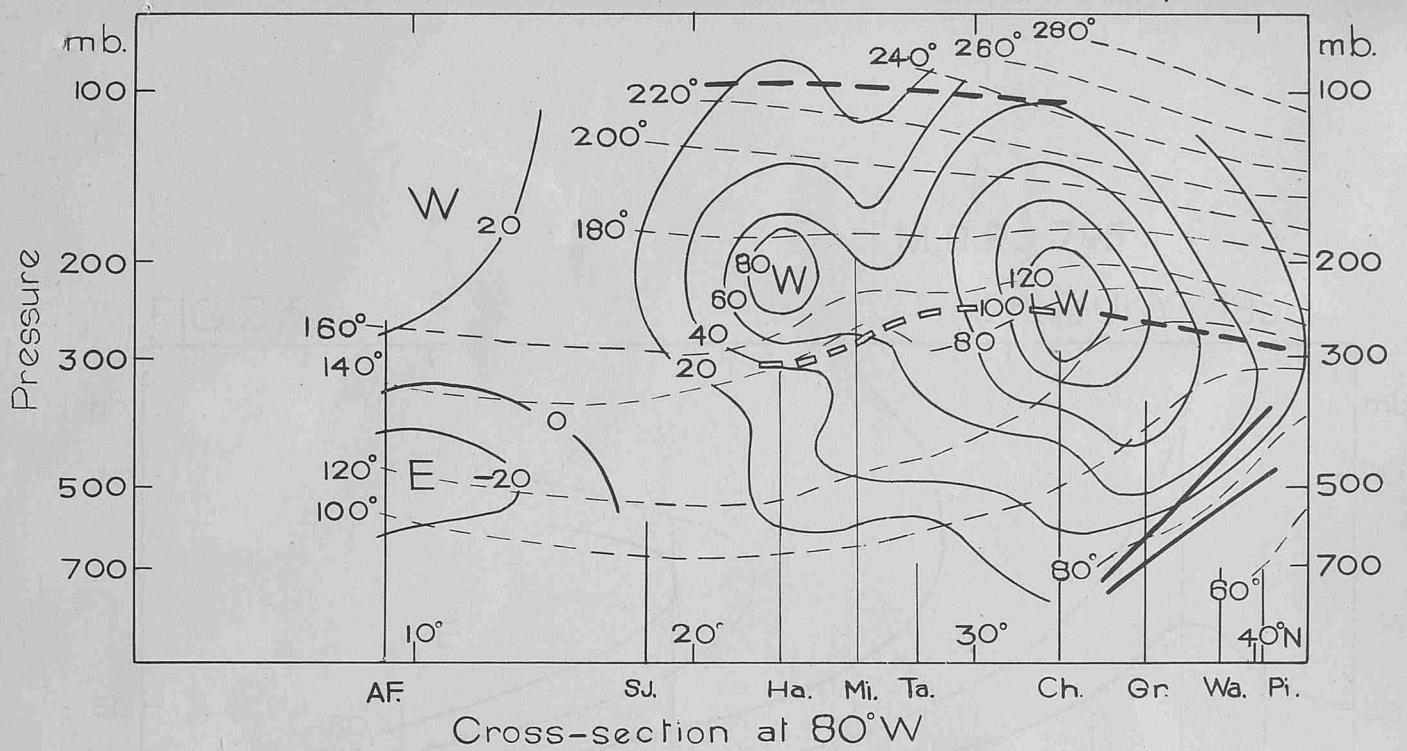
——— Isopleths of mean west wind speed (k.t.)

- - - - Isopleths of potential temperature (°F.)

— — Tropopause

FIG. 21

12th JANUARY, 1951



Cross-section at 80°W

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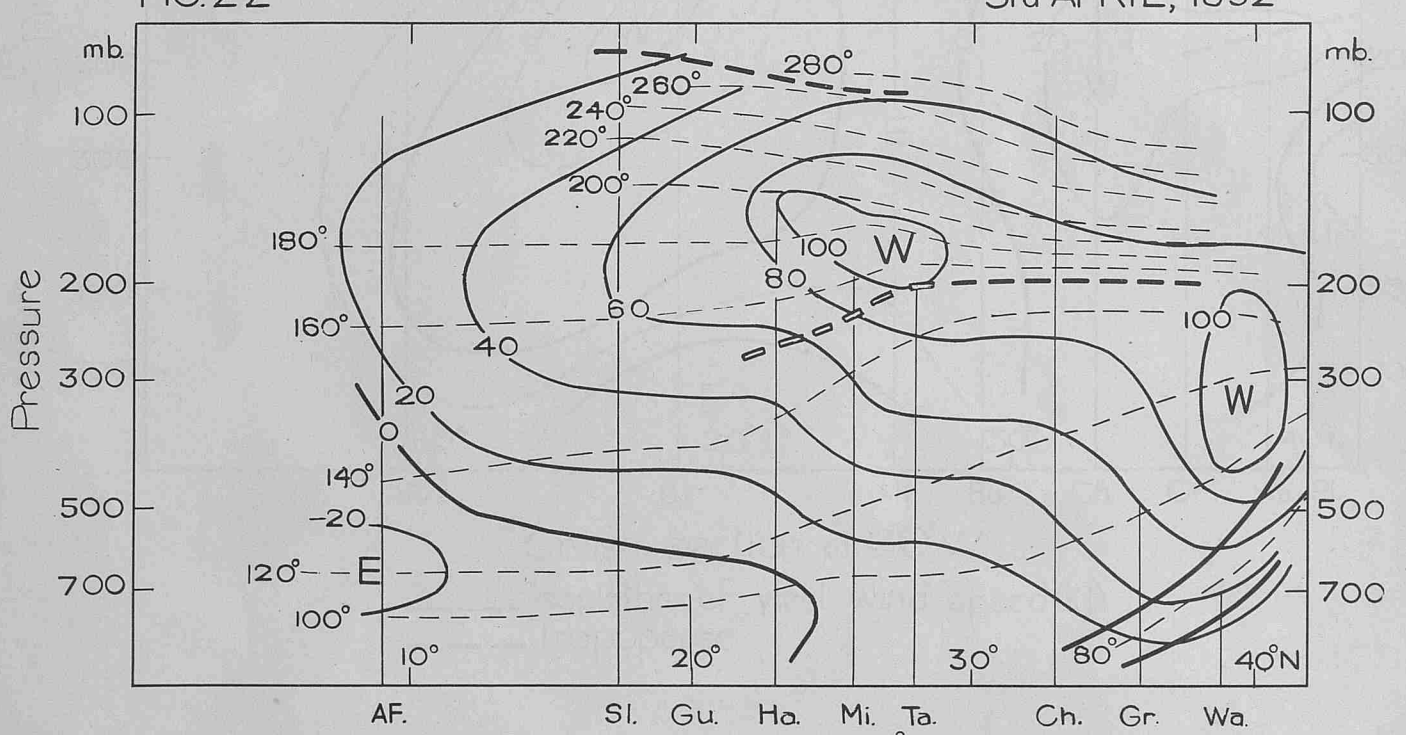
FIG. 22

3rd APRIL, 1952



Cross-section at 80°W

——— Isopleths of west wind speed (kt.)

- - - - - Isopleths of potential temperature (°F.)

— — — Tropopause ——— Front

— — — Continuation of tropopause surface, indicated by change of lapse rate—not a true tropopause

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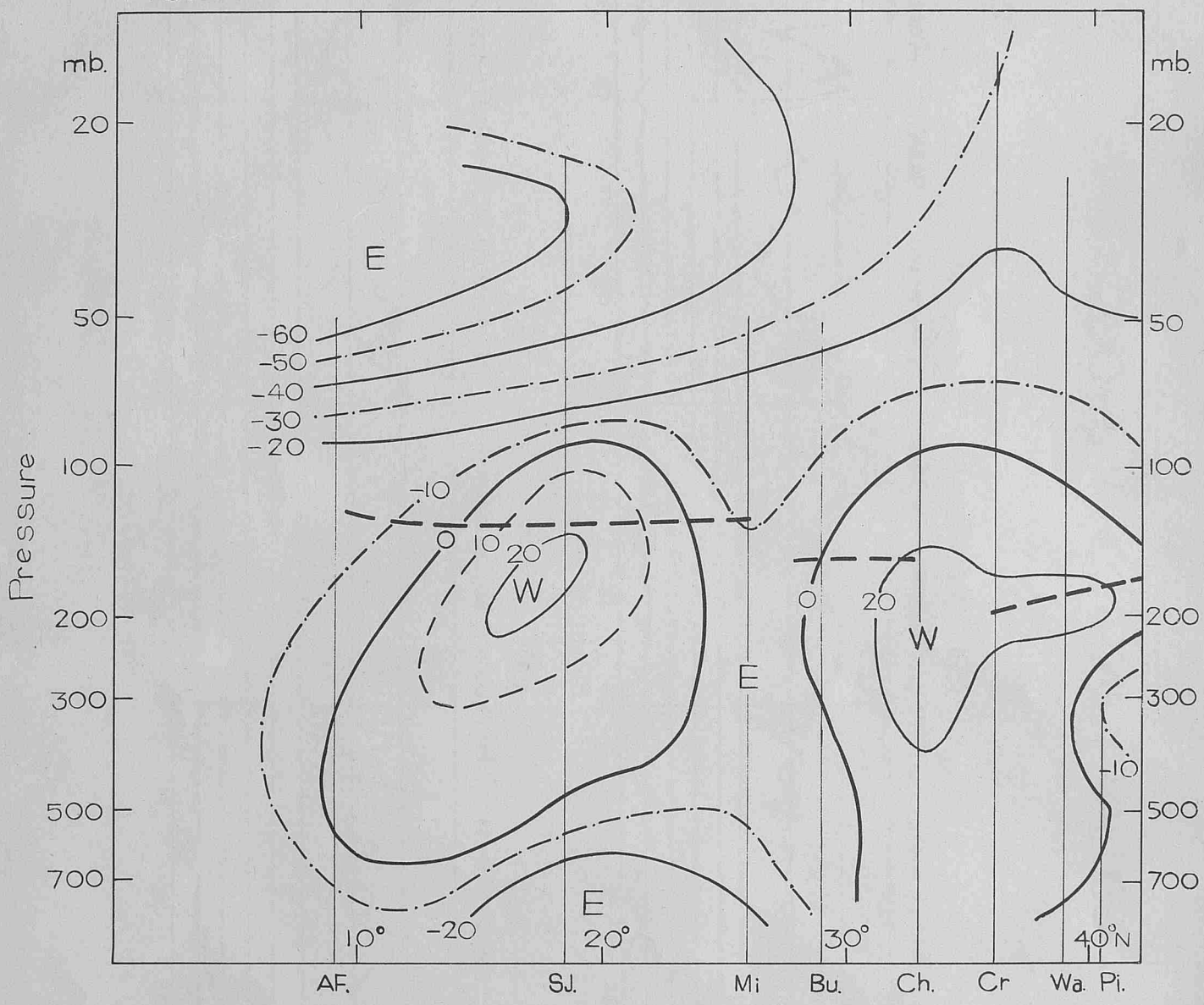
FIG. 23 12th JULY, 1952.



Cross-section at 80°W

——— Isopleths of west wind speed (kt.)

– – – Tropopause

200 mb. WINDS — DAILY VALUES

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FIG. 24

JANUARY 1951

(a) MIAMI (25°49'N 80°17'W) (b) DAKAR (14°40'N 17°26'W) (c) BAHREIN (26°16'N 50°37'E) (d) HILO (19°44'N 155°04'W) (e) ALBROOK (08°59'N 79°34'W) (f) NAIROBI (01°17'S 36°50'E)

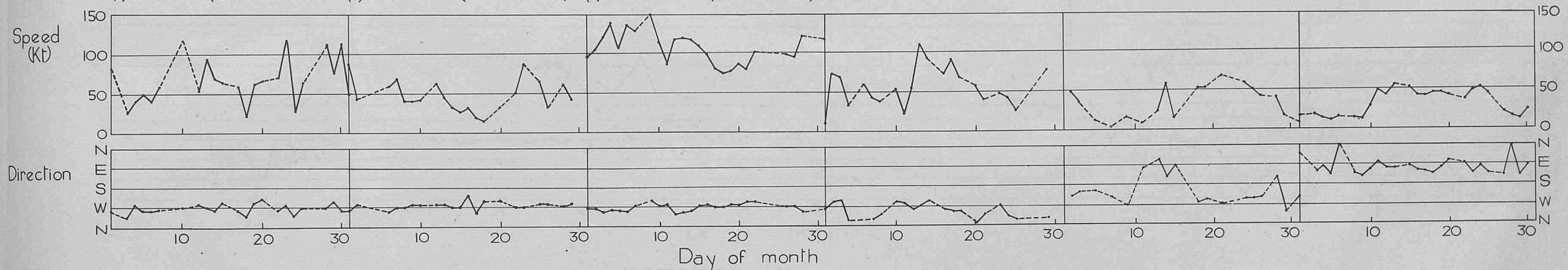
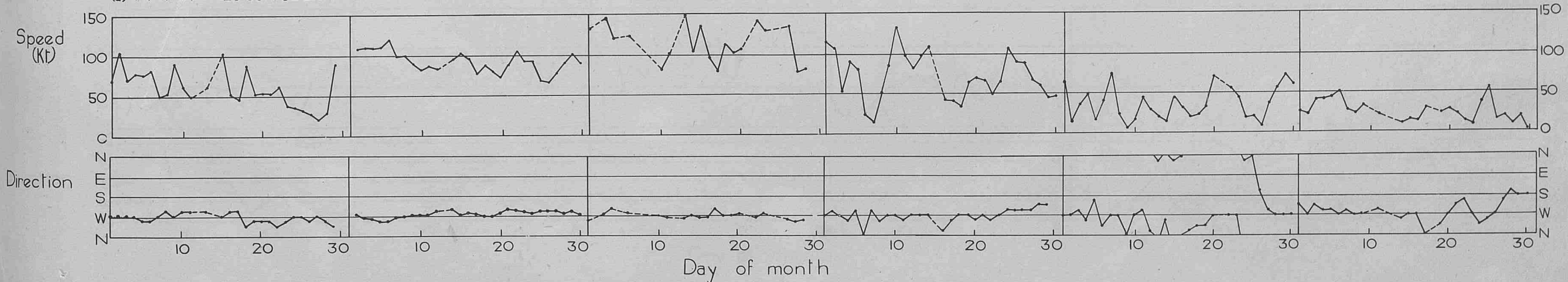


FIG. 25

APRIL 1951

(a) HAVANA (23°09'N 82°21'W) (b) BAHREIN (26°16'N 56°37'E) (c) TATENO (36°03'N 140°08'E) (d) HILO (19°44'N 155°04'W) (e) WAKE (19°18'N 166°37'E) (f) KWAJALEIN (08°43' 167°44'E)



200mb WINDS – DAILY VALUES

FIG. 26

JULY 1951 JULY 1951 JULY 1950

(a) ALBROOK (08°59'N 79°34'W) (b) ADEN (12°50'N 45°01'E) (c) NAIROBI (01°17'S 36°50'E) (d) HILO (19°44'N 155°04'W) (e) WAKE (19°18'N 166°37'E) (f) WAKE

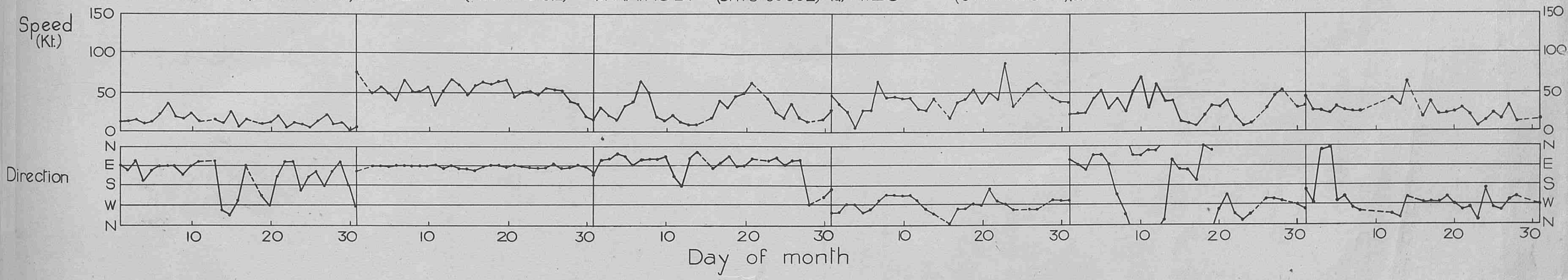
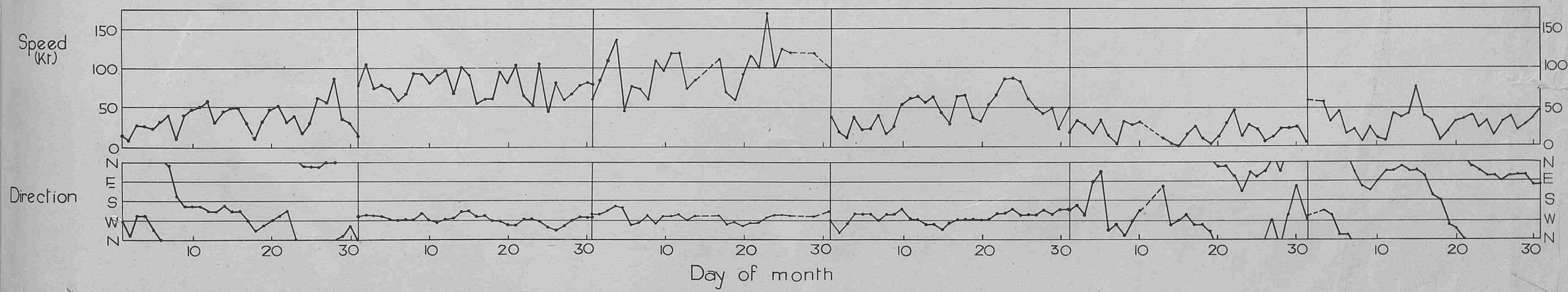


FIG. 27

OCTOBER 1951

(a) HAVANA (23°09'N 82°21'W) (b) HABBANIYA (33°22'N 43°34'E) (c) TATENO (36°03'N 140°08'E) (d) HILO (19°44'N 155°04'W) (e) ENIWETOK (11°21'N 162°21'E) (f) WAKE (19°18'N 166°37'E)



At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 797 by Mr. E.N. Lawrence A copy is available in the Library of the Meteorological Office."

M.R.P. 797
S.C.III/149
22nd April, 1953.

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The Effects of a Windbreak on the Speed and Direction of Wind

by

E. N. Lawrence (Meteorological Office)

SUMMARY

The following note describes the wind speed and direction to leeward of a windbreak, firstly in the horizontal plane at a height of about six feet (referred to as ground level), secondly in the vertical plane which passes at right angles through the centre of the barrier. The wind field at the end of the windbreak is also examined.

Two types of wind barrier are considered, one of which has a greater density than the other in the upper part of the screen.

The degree of shelter and the effect on wind direction were measured for a range of wind velocities and angles of incidence to the barrier. The wind speeds in the sheltered area have been calculated in terms of the speed of the "undisturbed" or "free" wind at the corresponding level for two types of atmospheric temperature lapse rate, which may be described as "large lapse rate" (referred to as unstable or lapse conditions) and "zero or small lapse rate" (referred to as stable or near neutral conditions).

INTRODUCTION

During the period August 1937 to January 1938, a series of observations was made at Manby, Lincolnshire, in order to explore the suitability of wind screens in reducing cross winds for the landing of aircraft (1). The development of this use of windbreaks did not materialize but the influence on wind of the two artificial barriers examined may be compared with that of natural shelterbelts of trees and shrubs of similar density, the use of which in agricultural work has commanded considerable attention in recent years. In conjunction with the Manby field experiment, wind tunnel investigations were carried out at South Farnborough in August 1936 and October 1938; the results therefrom are summarized in Appendix I.

METHOD

The windbreaks were approximately 50 feet in height, 1,600 feet long and 10 feet wide and are shown in the photographs of figures 1, 2, 3 and 4 (original windbreak) and figure 5 (modified windbreak). Details of the structure of the two barriers are shown in the diagrams of figures 7a and 7b. Figure 7 shows that the barrier consists of three parallel frames, the two external frames being similar. The modified barrier was constructed by adding slats (shown by shading in figure 7) to the upper parts of the frames.

/At

At ground level (6 feet), wind speeds were measured by cup anemometers mounted on tripods. The wind speed was obtained by averaging 2 or 3 runs, each run taking 3 minutes. Similarly, using a telescopic ladder (see figure 6), upper winds (30 feet to 55 feet above ground) were measured. For heights greater than those reached by the ladder (60 feet upwards), winds were measured with zero lift balloons. Flow conditions around the barrier were studied by means of balloons and smoke candles, the latter being observed every ten seconds for several minutes. Wind directions were estimated by means of light tape fixed at one end to a cup anemometer. These observations were aided by ground markings, consisting of white lines, parallel to the barrier. The mean direction was obtained from observations every five seconds over a period of one minute. Wind observations by cup anemometer were made at a height of 6 feet and at intervals of 5 feet from 30 feet to 55 feet, inclusive. In a horizontal plane, observations were made at 50 yard intervals up to 150 yards to leeward of the barrier and at 50 yard intervals parallel to the barrier, and at the "sheltered" end of the barrier on the leeward side, the observation network was denser and generally on a ten or twenty yard basis over the area in which the lines of equal "shelter" showed great curvature. Only one telescopic ladder was in use and observations were not simultaneous. It was not the general rule to "average" values for each point, although on occasions, two or more sets of observations were obtained for different values of the magnitude of the "free" wind. The speed and direction of the "free" wind were measured by a pressure-tube anemometer erected at the centre of the barrier-top. This instrument was suitably time-marked against a stop-watch, in order that observations of "free" wind could be compared with observations to leeward of the barrier.

Free winds below 9 m.p.h. were not sufficiently strong to give any definite results and the range of free velocities used in the calculations was 9 to 30 m.p.h. The wind directions were those corresponding to angles of incidence with the barrier of 0°, 15°, 30°, 45° and 60° for horizontal gradients and 15°, 30° and 45° for vertical gradients and 15° only for the examination of "end effect" of the wind field near the barrier end.

/CALCULATION

CALCULATION AND REPRESENTATION OF RESULTS

Let v be the pressure-tube anemometer reading and assume* that this is the velocity of the "free" wind at 50 feet.

Using Sutton's velocity profile formula (2)

$$v_z = v(z/50)^{\frac{n}{2-n}}$$

where v_z is the free wind at the height z. According to Sutton (3), when $n = 1/5$ (lapse conditions),

$$v_z = v(z/50)^{1/9}$$

and when $n = 1/4$ (near neutral conditions),

$$v_z = v(z/50)^{1/7}$$

Let u_z be the observed wind to leeward of the barrier at a height z and let $u_z = k\, v_z$.

Then $u_z = k\, v(z/50)^{1/9} = k'v$ say (lapse conditions)

and $u_z = k\, v(z/50)^{1/7} = k'v$ (near neutral conditions)

Now the available data give the values of k' at different points. The corresponding values of k are given by :-

$$k = k' (50/z)^{1/9} \quad \text{(lapse conditions)} \qquad \text{I}$$

and

$$k = k' (50/z)^{1/7} \quad \text{(near neutral conditions)} \qquad \text{II}$$

All the observations were made during the daytime and not normally under inversion conditions. If we assume a high lapse rate (lapse conditions), then, using equation I and substituting the values of k' for different points, we obtain a set of curves on which k =0.2, 0.4, 0.6 etc. Similarly, using equation II we obtain a set of curves for small lapse rates or near neutral conditions. These two sets of curves are a measure of the bounding positions of the lines of equal "shelter" or lines along which the ratio of the sheltered velocity to the free wind at the corresponding height is constant. On a horizontal section, these lines would be lines of equal velocity.

---

*The assumption that v is the velocity in the open at 50 feet appears to be reasonable, in view of the vertical cross section obtained by Bates and Stoeckeler of the United States Forest Service (4). For a half-solid, sixteen feet barrier, the velocity at a short distance above the top was found to be near to 100 per cent of the velocity at a corresponding height in the open. In the following results, the degree of shelter would be over-estimated, if the velocity at the barrier top were under-estimated or if the height in the free air corresponding to this velocity were under-estimated. For example, if the pressure tube anemometer were actually recording 110 per cent of v_{50},

$$v = 11\, v_{50}/10$$

$$= 11\, v_z(50/z)^{1/9}/10 \quad \text{or} \quad 11\, v_z(50/z)^{1/7}/10$$

using Sutton's formula (2).

True "k" = k_T say $= u_z/v_z$

$$= 11\, u_z(50/z)^{1/9}/10v \quad \text{or} \quad 11\, u_z(50/z)^{1/7}/10v$$

$$= 11\, k/10$$

/Thus

Thus, if v were actually 110 per cent of the free wind at 50 feet, all values of k would have to be increased by 10 per cent.

RESULTS

The results are shown in graphical form, as follows (the arrows denoting the directions of general flow):-

Original windbreak (see figs. 7a and 7b)

Figs. 9, 10, 11, 12, 13 (Velocity contours at ground level for angles of incidence of 0°, 15°, 30°, 45°, 60°)

Figs. 14, 15, 16 (Shelter contours behind centre of screen for angles of incidence 15°, 30°, 45°)

Fig. 17 (Shelter distribution near end of screen at 60 ft. from screen, wind at 15°)

Fig. 18 (Large eddies observed behind centre of screen with winds at large angle to screen)

Modified windbreak (see figs. 7a and 7b)

Figs. 19, 20, 21, 22 (Velocity contours at ground level for angles of incidence 15°, 30°, 45°, 60°)

Figs. 23, 24 (Shelter contours behind centre of screen for angles of incidence 15°, 30°)

Fig. 25 (Wind roses at ground level at points 3H and $4\frac{1}{2}$H behind centre of screen - where H is the height of the barrier and angle of incidence is 15°)

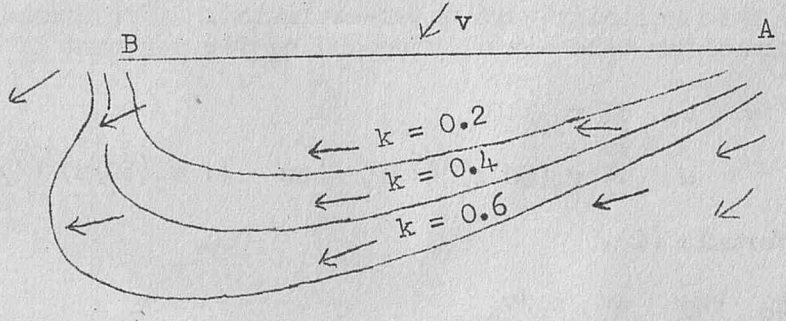
Note: The angle of incidence of the wind to the barrier (I) includes the range of wind directions given by I $\pm$ 5 degrees.

CONCLUSIONS

As the observations (1) were carried out with a windbreak of height 50 feet and length 500 yards approximately, the conclusions apply primarily to a barrier with dimensions of this order. The same results cannot be expected with a barrier in which the aspect ratio (i.e. ratio of length to height) is much smaller.

Original windbreak

(i) The general form of the contours at ground level is similar for all wind directions and is illustrated below. The shielded



area decreases as the angle of incidence of the wind to the barrier increases, and the direction of the wind is altered so that the flow becomes approximately parallel to the screen and is smooth.

/The

The extent of the area in which the flow is parallel to the screen varies greatly with the angle of the wind. At B the flow is not smooth.

(ii) The vertical cross-sections through the centre of the screen and at right angles to it are very similar for different wind directions and at a distance of 1H from the screen are almost identical. Up to a height of about 0.7H, there is a tendency for the flow to become parallel to the windbreak but above this height, the direction of flow is approximately that of the "undisturbed" wind.

(iii) Eddying conditions

At distances downwind greater than $4\frac{1}{2}$ H from the windbreak, the wind direction at "ground level" was very variable for angles of incidence smaller than 20°. With an angle of incidence of about 25°, at a distance from the screen between $4\frac{1}{2}$ H and 9 H, the flow near the ground was generally along the direction of the barrier but gusts were observed towards and less frequently away from the screen. With a wind of zero mean angle of incidence (i.e. perpendicular to barrier), the flow near the ground at distances greater than about $4\frac{1}{2}$ H from the barrier consisted of a series of weak gusts in almost every direction, the average velocity at 6 H being approximately .25 V 50.

With a wind of zero mean angle of incidence, candle smoke released from the top of the barrier showed a steady stream with small eddies rising to about 1.2 H above the ground at 3 H from the barrier. Almost all the smoke reached the ground at distances greater than 6 H from the barrier. Occasionally some smoke was brought down to the ground by large, slow eddies at distances as near as $1\frac{1}{2}$ H from the barrier. The general flow is illustrated in fig. 18 and a photographic observation of an eddy is shown in fig. 8. Occasionally a balloon with no lift and floating a few feet above the ground to leeward of the barrier would be caught in a rapid updraught until it was carried away by the main stream over the top of the windbreak.

Modified windbreak

(i) The contours near the ground have the same general form as with the original barrier but the shielded area is rather less than the previous area for angles of incidence of 15° and 30° and slightly greater for angles of incidence of 45° and 60°.

(ii) Similar contours are a few feet higher in the vertical plane than the corresponding contours with the original barrier up to a distance of about 6 H from the windbreak.

(iii) The eddying conditions in the sheltered region are more disturbed than with the original barrier. As in the case of the latter, with a zero mean angle of incidence, most of the smoke reached the ground at distances of more than 6 H from the barrier. Large eddies which carried the smoke down to the ground were more frequent and more violent than those observed with the original barrier.

With angles of incidence of wind between 0 and 20°, the wind direction on the ground at distances of 3.6 H to 6 H was very variable with fairly strong frequent gusts towards the barrier. Smoke released in this area was carried towards the barrier and generally at about 3.6 H distance rose rapidly to a height of 1.2 H. These gusts did not often penetrate to within a distance of 3 H from the windbreak. The wind roses of fig. 24 illustrate this. At a distance of 3 H, the most frequent wind direction is approximately parallel to the screen.

/Both windbreaks

Both windbreaks

(i) The effect of lapse rate differences (for positive lapse rates) is very small except below a height of approximately $\frac{1}{2}$ H and at distances from the barrier between about 2 H and 8 H, especially for angles of incidence of about 30° or less.

(ii) The effective reduction of wind is normally greater for unstable than for stable lapse rates up to a height of H.

(iii) The shelter or velocity contours of these artificial barriers may be compared with those obtained by Nageli (5) for a gap in a hedge. As might be expected, the shelter contour gradient becomes much greater near the gap and presumably the narrower the gap in the barrier, the greater will be this gradient. Also, the shelter near a gap does not extend so far beyond the "ends" of the hedge as with the single end of the artificial barriers described. The general run of the contours is otherwise similar.

APPENDIX I

A NOTE ON WIND TUNNEL EXPERIMENTS

These experiments were originally devised to obtain by means of a windbreak a sufficiently extensive area of shelter without severe velocity gradients (that is, variations of velocity with distance) or eddying conditions in the boundary zone. A solid screen was known to cause a very disturbed eddy and to give only a relatively small extent of protected area. The experiments (1936) (6) were therefore directed towards the development of a perforated barrier.

In the Manby (full-scale) experiments, flat slats were used instead of the gauze of the model tests. As the drag on flat plates is greater, this was compensated for by using a smaller windbreak density in the full scale experiments. In spite of this, the full scale experiments did not produce proportionally so large a sheltered area as the models.

Further model experiments were carried out in 1938 (7), using copper ribbon soldered to steel brackets and arranged to give the same density as in the full scale experiments. The sheltered area was considerably less than in the earlier model tests but again greater than on the full scale. The differences between the two sets of wind tunnel experiments were attributed to differences in screen design and to tunnel boundary effects. The differences between the 1938 model tests and the full-scale experiments increased with distance from the barrier and were accounted for partly by the constriction of the main stream by the tunnel walls and partly by the different methods of wind velocity measurement. On the full scale, the mean absolute velocity regardless of direction was measured by a cup anemometer. In the model tests the velocity recorded by the pitot-static tube in approximately the mean wind direction would be lower than that shown by a cup anemometer in regions where frequent changes in wind direction occurred.

Concerning the effects of scale and aspect ratio, the available evidence on the drag of plates (8) of infinite aspect ratio shows no scale effect above a Reynolds number (based on width normal to wind) of about 4,000. Below this value the drag coefficient decreases to a minimum of 85 per cent of its steady value at an Re of 1,500. The Re of the individual strip in the model tests was about 3,000 at full stream velocity. No evidence was known as to the scale effect on the dimensions of the wake of a flat plate or lattice girder, but it was assumed, on the basis of drag measurements (8), that the scale effect would be small and would not account for a larger sheltered area at the lower Reynolds number.

To examine end effects in the tunnel, it was found necessary and acceptable to use a much smaller aspect ratio (i.e. 11.9).

/REFERENCES

REFERENCES

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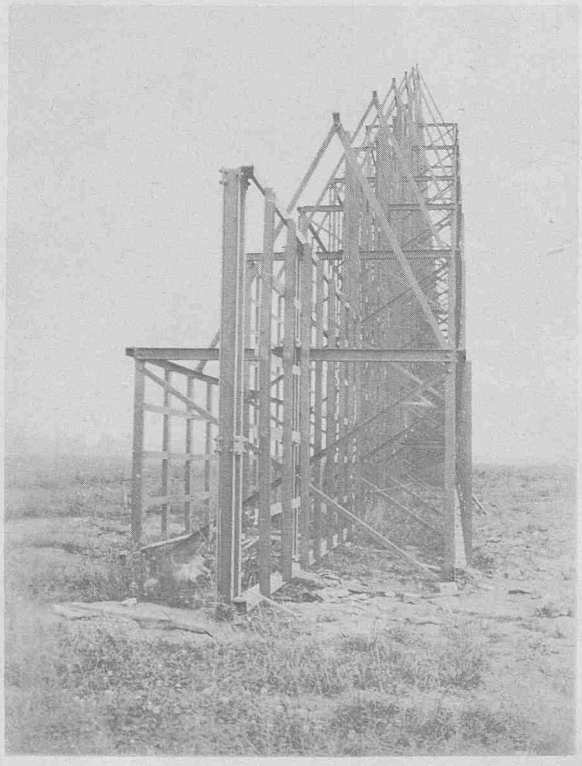


Fig. 3

Endview of the original windbreak.

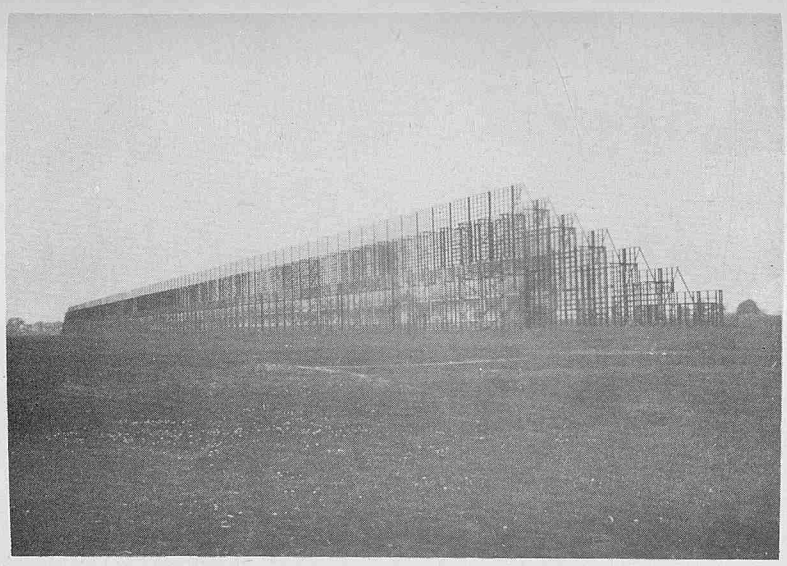


Fig. 1

General view of the original windbreak.

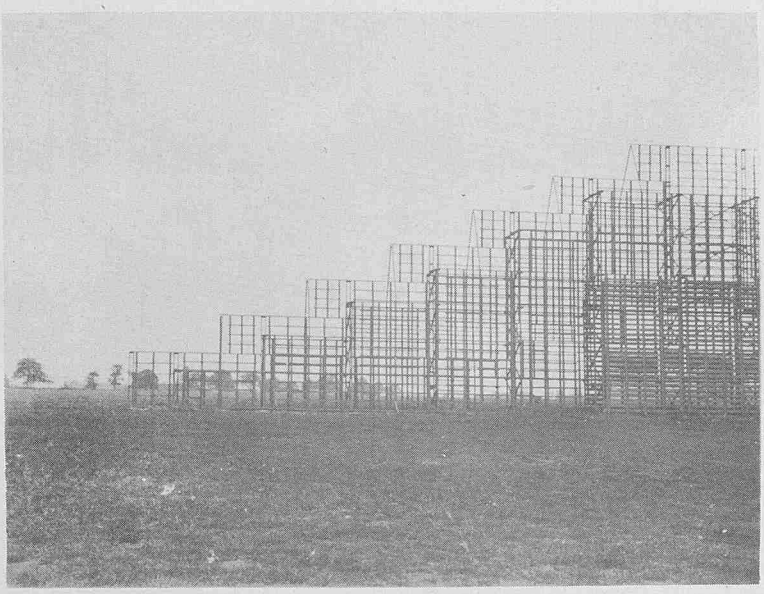


Fig. 2

The original wirdbreak showing the taper and the gradual reduction in density (or blockage ratio) towards the end.

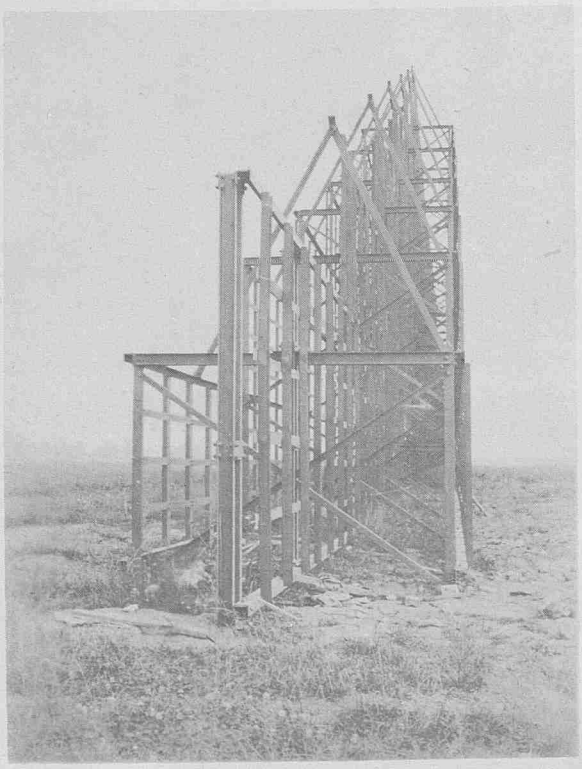


Fig. 3

Endview of the original windbreak.

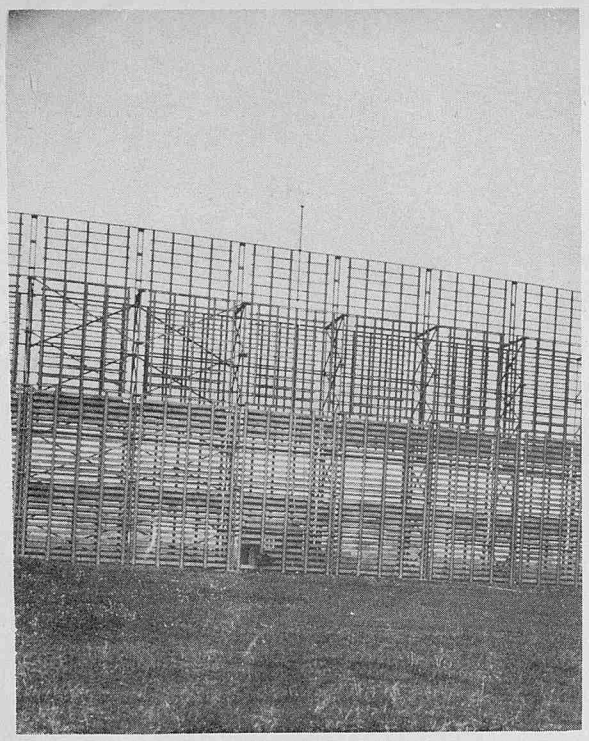


Fig. 4

The central portion of the original windbreak showing pressure tube anemometer.

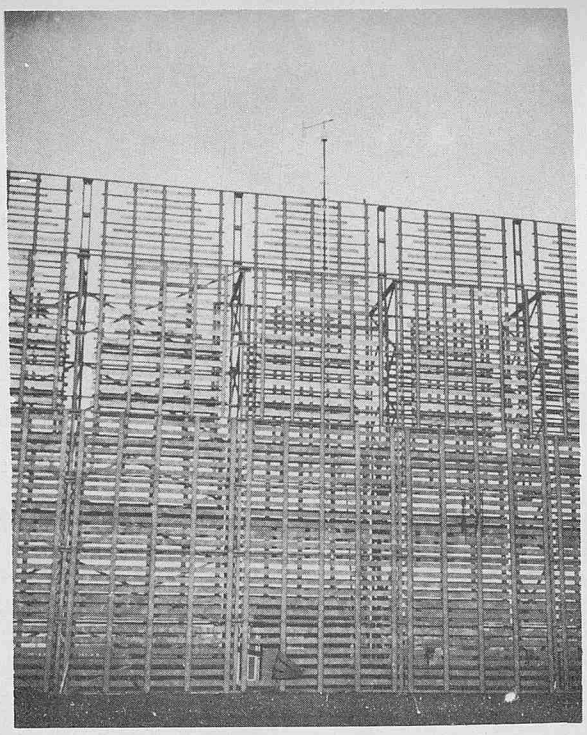


Fig. 5

The central portion of the modified windbreak.

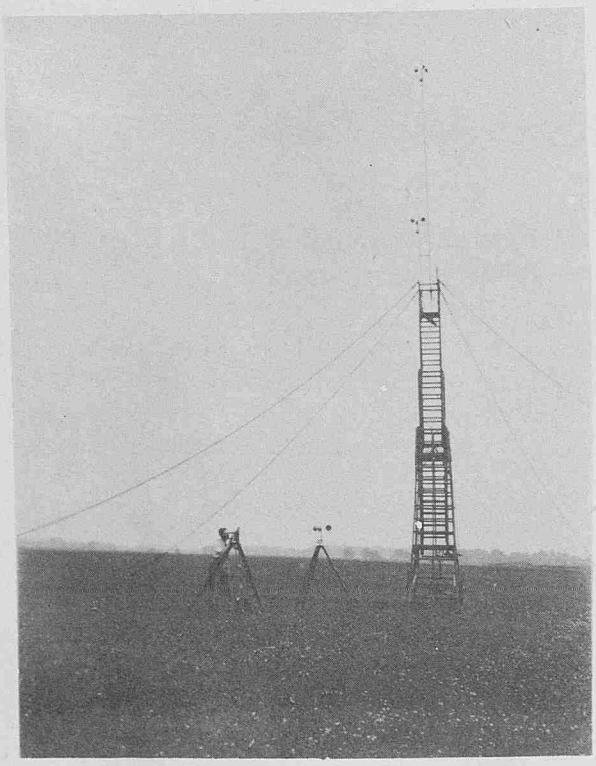


Fig. 6

The telescopic ladder with mounted cup anemometers used to measure winds up to a height of 60 feet. (Note the use of the theodolite to take readings of the anemometers.)

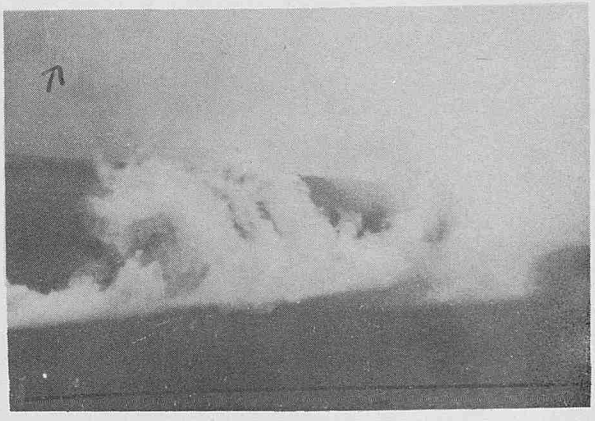


Fig. 8

A smoke eddy (Source at A) with the wind at an angle of incidence of 15° (from the left) to line of windbreak (BC).

Fig. 7a

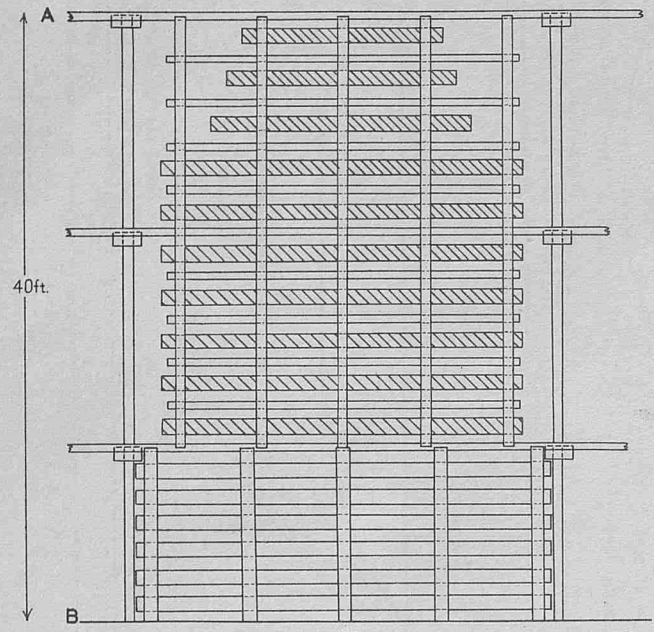
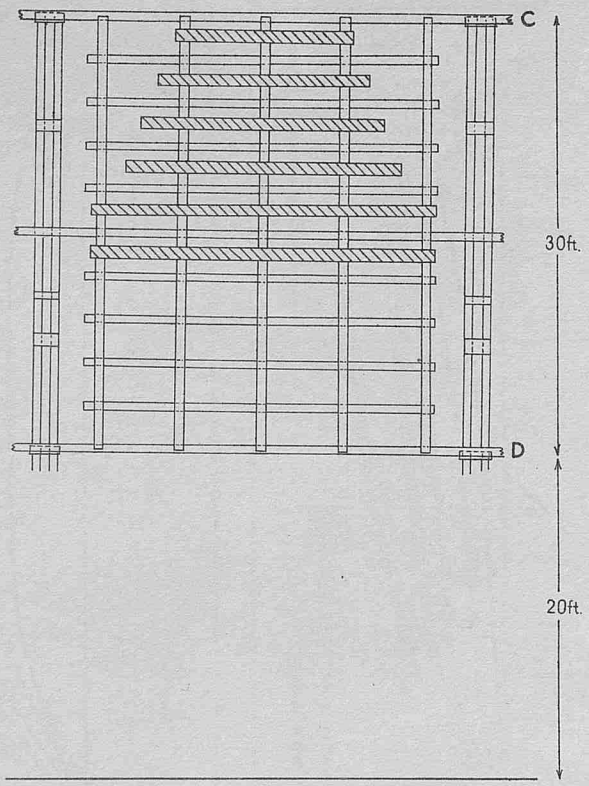
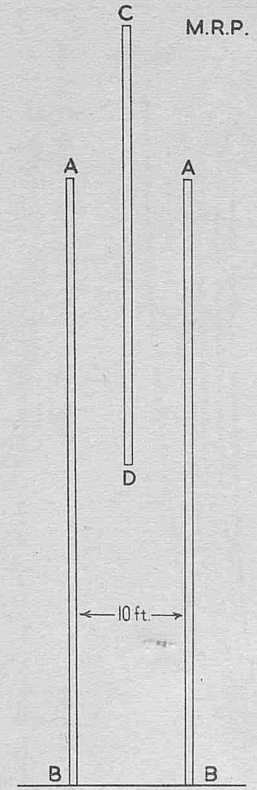


Fig. 7b



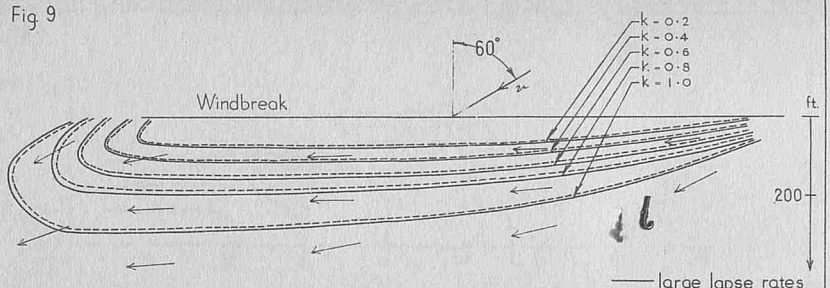


M.R.P. 797

A section of the length of the external frames (fig. 7a) and the internal frame (fig. 7b) of the windbreak, showing the additional slats (shaded) of the modified windbreak.

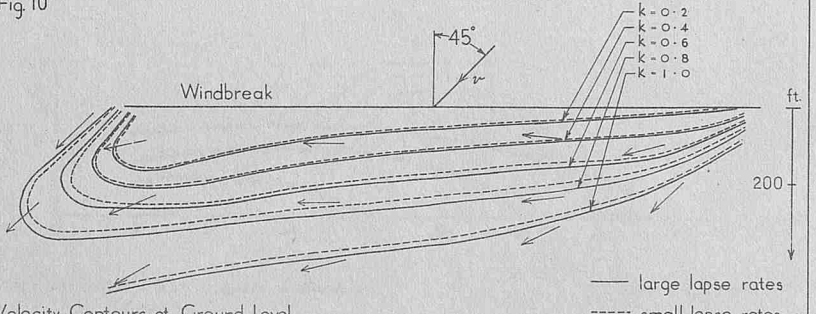
M.R.P. 797

Fig. 9



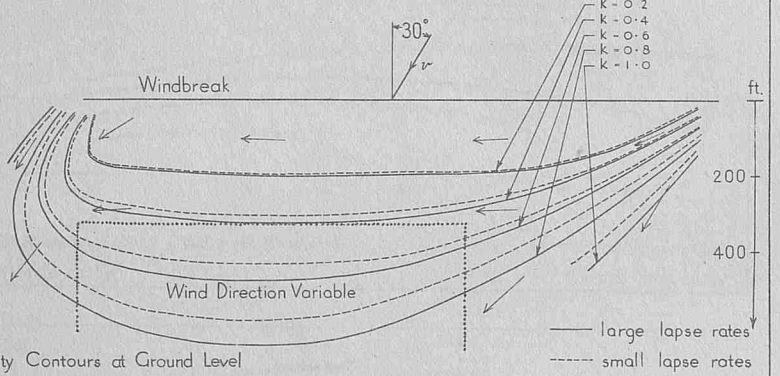
Velocity Contours at Ground Level
Length of Windbreak - 1600 ft. approx.

Fig. 10



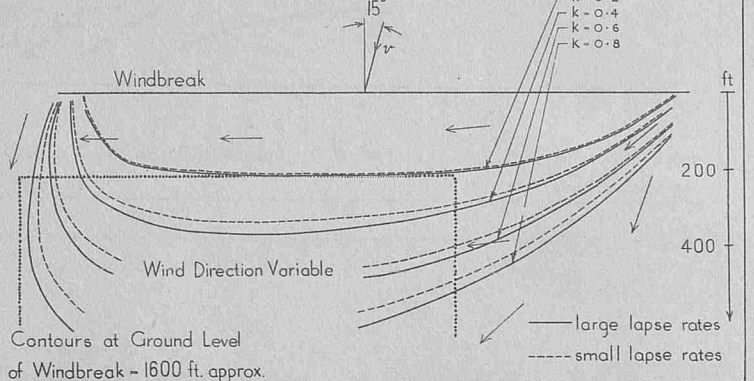
Velocity Contours at Ground Level
Length of Windbreak - 1600 ft. approx.

Fig. 11



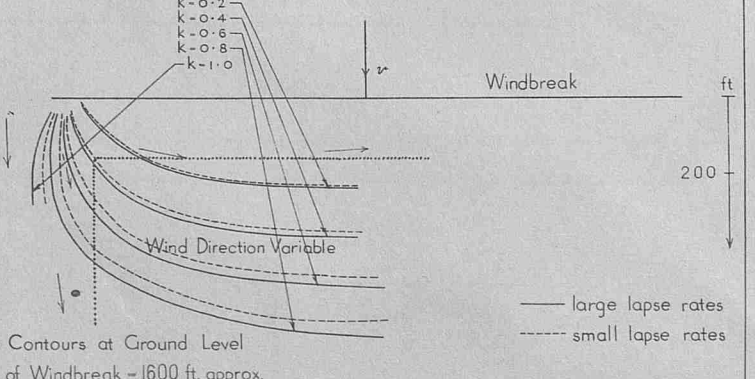
Velocity Contours at Ground Level
Length of Windbreak - 1600 ft. approx.

Fig. 12



Velocity Contours at Ground Level
Length of Windbreak - 1600 ft. approx.

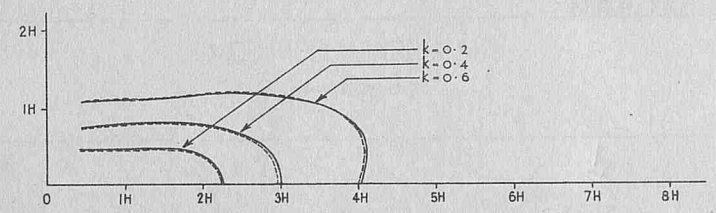
Fig. 13



Velocity Contours at Ground Level
Length of Windbreak - 1600 ft. approx.

ORIGINAL WINDBREAK

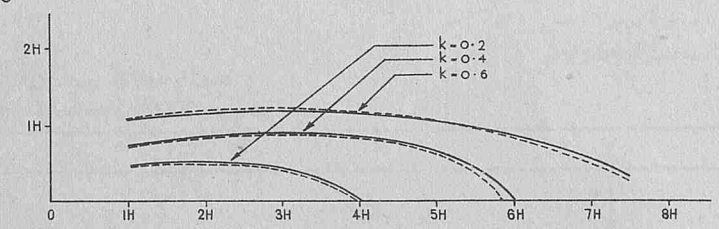
Fig. 14



Shelter Distribution behind Centre of Windbreak
Angle of Incidence 45°

—— large lapse rates
----- small lapse rates

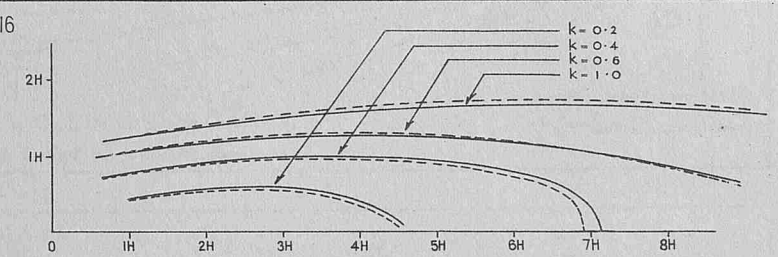
Fig. 15



Shelter Distribution behind Centre of Windbreak
Angle of Incidence 30°

—— large lapse rates
----- small lapse rates

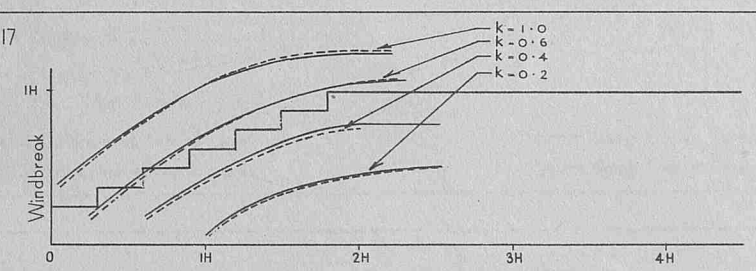
Fig. 16



Shelter Distribution behind Centre of Windbreak
Angle of Incidence 15°

—— large lapse rates
----- small lapse rates

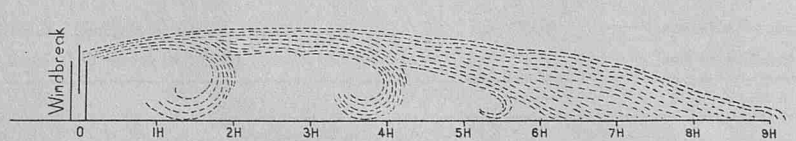
Fig. 17



Shelter Distribution Near End of Windbreak at 60 feet
Angle of Incidence 15°

—— large lapse rates
----- small lapse rates

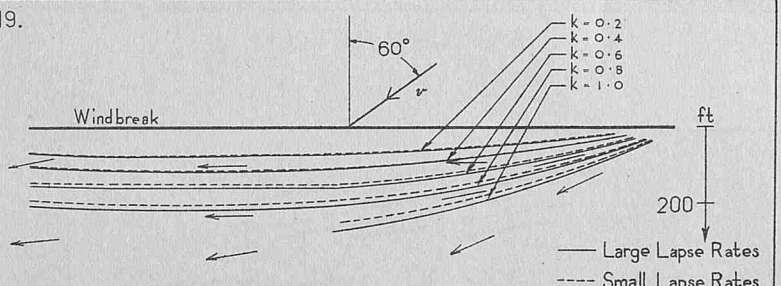
Fig. 18



Large Eddies Observed Behind Centre of Windbreak
Small Angles of Incidence

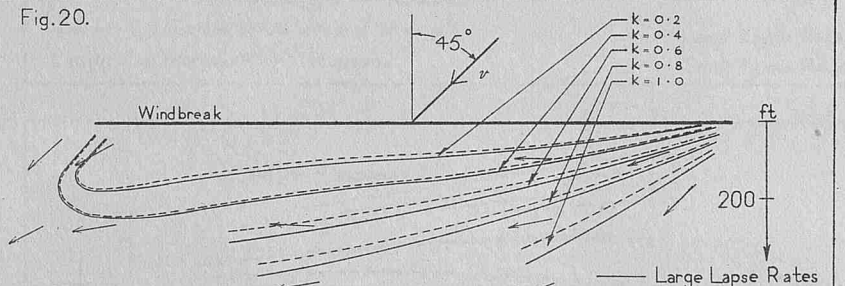
MODIFIED WINDBREAK
(see Fig. 8.)

Fig. 19.



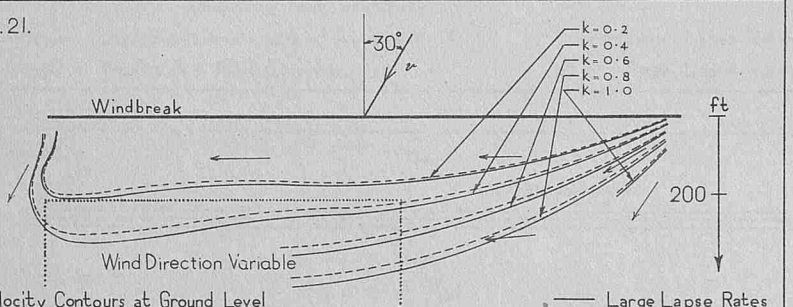
Velocity Contours at Ground Level
Length of Windbreak = 1600ft. approx.

Fig. 20.



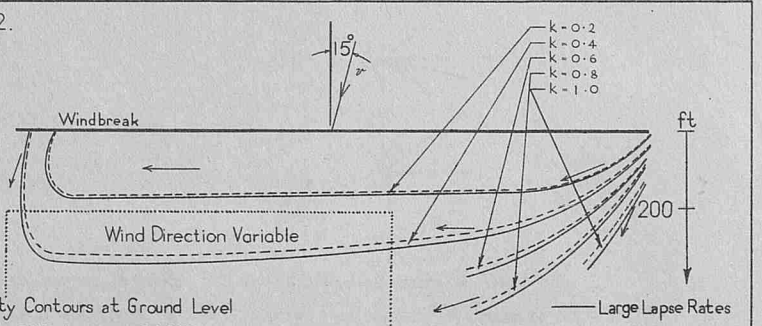
Velocity Contours at Ground Level
Length of Windbreak = 1600ft. approx.

Fig. 21.



Velocity Contours at Ground Level
Length of Windbreak = 1600ft. approx.

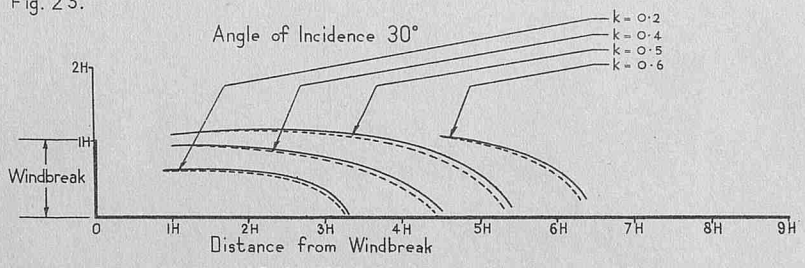
Fig. 22.



Velocity Contours at Ground Level
Length of Windbreak = 1600ft. approx.

MODIFIED WINDBREAK

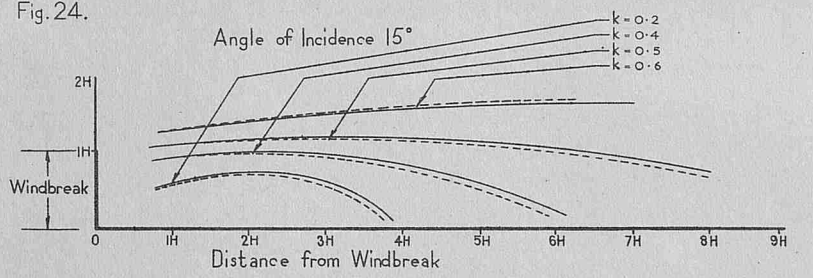
Fig. 23.



Shelter Distribution behind Centre of Windbreak
Length of Windbreak = 1600 ft. approx.

—— Large Lapse Rates
------ Small Lapse Rates

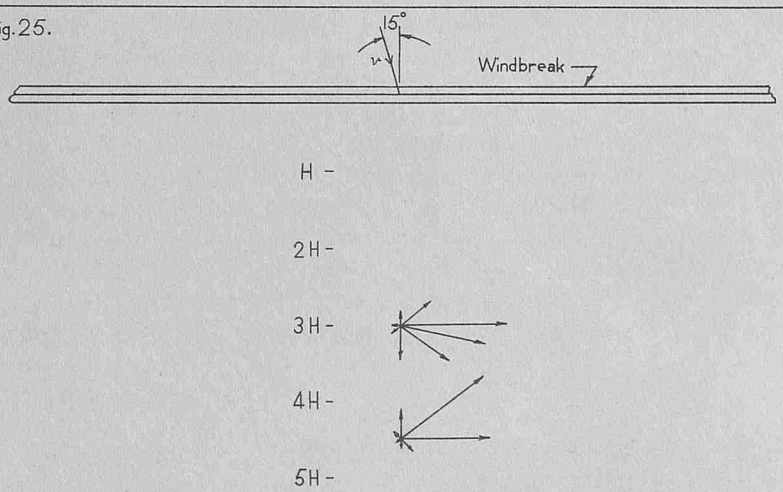
Fig. 24.



Shelter Distribution behind Centre of Windbreak
Length of Windbreak = 1600 ft. approx.

—— Large Lapse Rates
------ Small Lapse Rates

Fig. 25.



Direction roses at points 3 H and 4½H behind centre of Windbreak
Length of Windbreak = 1600 ft. approx. Height of Windbreak H = 50 ft.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London),
M.R.P. No. 799 by Mr. J.S. Sawyer and Mr. F.E. Dinsdale
A copy is available in the Library of the Meteorological Office."

M.R.P. 799
S.C.II/143
23rd April, 1953.

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Cloud in relation to active warm fronts near Bircham Newton during the period April, 1942 to April, 1946

by
J.S. Sawyer and F.E. Dinsdale (Meteorological Office)

1. Introduction

In M.R.P. 747 Matthewman reported an investigation of the cloud structure of warm and quasi-stationary fronts which occurred near Bircham Newton during the three months January to March, 1942. Matthewman found that the cloud structure was not related closely with any of the synoptic parameters which he selected to represent the activity of the front. As his sample of fronts included a number of feeble and inactive fronts, it was thought possible some more definite relationships might have emerged if attention has been confined to active warm fronts. However this has not proved to be so, and the present paper reports the results of an analysis of the cloud structure of a series of active warm fronts; the analysis was restricted to fronts which had a rain belt at least 50 miles wide.

As a measure of the extent of cloud at the front Matthewman adopted a classification based essentially on the proportion of the vertical column between the ground and 400 mb. which was occupied by cloud. The cloud class number so determined tended naturally to increase towards the front, and it was thought that any relations between the class number and the other synoptic parameters might have been obscured thereby. To avoid this difficulty a classification of "cloudiness" was attempted on various other bases, but the essential results of the investigation were unaffected.

2. The Data and Technique

During the period 1942 - 1946 aircraft ascents were made at least twice daily at Bircham Newton or Langham. These ascents ceased in May, 1946 and it was not therefore possible to extend the period of investigation into the post-war years.

The British Daily Weather Reports were examined for the 4 years and 76 cases were found where the surface warm front was reasonably well determined and where the aircraft temperature sounding showed clearly both upper and lower boundaries of the frontal zone below or at the 400 mb. level. The ascents were seldom synchronous with the surface chart, but did not usually differ by more than one hour.

In each case a vertical cross-section was drawn along the normal from Bircham Newton to the surface front utilising any other sounding, whether British or foreign, near this normal and not differing by more than 2 or 3 hours from the time of the chart. On each cross-section were plotted the positions of the upper and lower boundaries of the frontal zone at each ascent together with the reported cloud.

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In analysing the soundings to select the frontal zone only broad thermal considerations were employed. No attention was paid to humidities. In some cases a small adjustment was made to the position of the surface front or of an ascent in order to correct for a time discrepancy where this gave a distorted picture.

Measurements of surface winds and speeds of fronts were made on the British Daily Weather Reports. The upper winds were taken from Downham Market, interpolating in time and space where necessary. The Downham Market wind soundings however seldom differed by more than one hour from the Bircham Newton ascents.

It was found that in 7 out of the 76 cases the warm air could not be satisfactorily identified at the surface, and the surface front marked on the charts could not therefore be taken as the lowest point on the upper boundary of the front. In tests involving the distance of Bircham Newton from the surface front these 7 cases were omitted for the sake of uniformity.

3. Cloud Classification

Matthewman classified the vertical distribution of cloud reported by the Bircham Newton aircraft into 6 groups which for clarity are given below:-

Class 1. No cloud.

Class 2. Cloud below or near the base of the frontal zone only (except cirrus or cirrustratus).

Class 3. One or more thinnish layers of cloud above or near the top of the frontal zone. Here thinnish implies not more than 150 mb. deep.

Class 4. Thick cloud extending above the frontal zone but with one or more layers or lanes of clear air at least 150 mb. deep between the ground and 400 mb. Here thick implies more than 150 mb. deep.

Class 5. Continuous or almost continuous cloud extending from near the ground to near or above 400 mb. but with one or more layers or lanes of clear air not more than 150 mb. deep.

Class 6. Continuous cloud extending from near the ground to near or above 400 mb.

The 76 cases were first classified according to this system and relationships determined between the class numbers (referred to hereafter as the C number) and the various parameters suggested in the original paper. Tables I to III show that the C number is very much dependent upon the distance of the ascent from the surface front. This of course is no more than synoptic experience indicates, but as one of the objects of the paper was to determine what simple measurements a forecaster might make in order to estimate the general cloudiness of a warm front, an attempt was made to obtain a classification which was not dependent on the distance from the surface front. For this purpose attention was confined to cloud in or above the frontal zone. Nevertheless the association of class number with distance from the surface front was very little reduced. One such classification is given below:-

X Classification

1. No cloud in or above the frontal zone except cirrus above 400 mb.

2. Cloud occupying less than 25% of the space between the lower limit and 400 mb.

3. Cloud occupying 25% to 75% of the space between the lower limit and 400 mb.

/4.

4. Cloud occupying more than 75% but not all of the space between the lower limit and 400 mb.

5. Cloud continuous from the lower boundary to 400 mb.

The correlation coefficients between distance from the surface front and the class numbers C and X are both about - 0.6.

Finally, taking the X classification, the regression equation of the X number upon distance from the front was calculated:- X = 4.0 - .005 D (where D is distance in miles and 69 cases were included). The departures of the X numbers from the values calculated from the regression equation were obtained to form a new classification Y, and correlation coefficients between Y and the various parameters determined.

It is to be noted that in none of the 76 cases was there a cloudless ascent at Bircham Newton, and hence C1 has been omitted from the following tables.

4. Results

(I) Relation of cloud at Bircham Newton to the cloud at simultaneous soundings

There were in all 41 other aircraft ascents, mainly foreign, which were used in the cross-sections. These were also given a C number and then compared with the Bircham Newton ascents. The following table sets out the number of cases in which cloud class number was greater than, equal to, or less than that at Bircham Newton for simultaneous observations towards and away from the warm front.

Table I

Relation of the cloud at Bircham Newton to the cloud at simultaneous soundings

	C number less than Bircham Newton	C number equal to Bircham Newton	C number greater than Bircham Newton
Sounding towards warm front	0	0	3
Sounding away from warm front	27	11	0

This table confirms that in individual cases the C number increases towards the front.

(II) Relation of cloud at Bircham Newton to the distance from Bircham Newton of the midpoint of the frontal zone at the ground

The distances ranged from 20 miles to 450 miles, and the mean distance decreased steadily with increasing class number, ranging from 305 miles in C2 to 66 miles in C6. There was considerable scatter about the mean in each class. Table II shows the C number tabulated according to 3 ranges of distance from the surface front (69 cases). It shows the tendency for the class number to increase towards the front.

Table II

No. of cases of specified cloud structures within stated ranges of the distance from the front

Distance of Bircham Newton from surface front	C number 2	3	4	5	6	All
Less than 100 miles	0	4	6	9	4	23
100 - 300 miles	4	12	13	3	1	33
More than 300 miles	9	3	1	0	0	13

(III) Relation of cloud at Bircham Newton to the pressure at (a) the upper (warm) boundary and (b) the lower (cold) boundary of the frontal zone

As might have been expected from the previous paragraph the means of both (a) the pressure at the upper boundary, and (b) the pressure at the lower boundary increased steadily with increasing C number. Again there was considerable scatter about the mean in each class.

Table IIIa

Relation of cloud at Bircham Newton to the pressure at the upper boundary

	C number (76 cases)				
	2	3	4	5	6
Mean pressure in mb.	525	601	665	685	760

Table IIIb

Relation of cloud at Bircham Newton to the pressure at the lower boundary

	C number (76 cases)				
	2	3	4	5	6
Mean pressure in mb.	747	826	869	890	940

(IV) Relation of cloud at Bircham Newton to the slope of the warm boundary of the frontal zone

As it seems probable that there is usually some variation in the slope of the frontal zone at both warm and cold boundaries, it is necessary to specify exactly what was taken as the slope for the purpose of comparison between classes. Here it was defined simply as the ratio of the height of the warm boundary at Bircham Newton to the distance of Bircham Newton from the surface front. In the seven cases where the warm air did not reach the ground, the slope was measured to the point vertically above the surface front that the warm boundary was considered to have reached.

It will be seen from Table IV that there is some increase of slope with increasing C number, but the scatter about the mean in all cases was large. In order to determine if any relation existed between the "cloudiness" and the slope of the front when the effect of distance from the front was eliminated, the Y number was correlated with frontal slope; the resulting correlation coefficient was negligible.

Table IV

Relation of cloud at Bircham Newton to the slope of the warm boundary

	C number (76 cases)				
	2	3	4	5	6
Mean slope	.010	.012	.013	.017	.014

The overall mean slope was .013 - slightly greater than the conventional value of 1 in 100 often quoted as typical for warm fronts.

/(V)

(V) Relation of cloud at Bircham Newton to the pressure thickness of the frontal zone

The average of all 76 cases was 210 mb., and the table shows that there was some decrease of the mean pressure thickness with increasing class number. This is probably due to the tendency for the frontal zone to become somewhat deeper or more diffuse away from the surface front. There was again a wide scatter in each class e.g. from 50 mb. to 350 mb. in C5.
The correlation with Y number was zero.

Table V

Relation of cloud at Bircham Newton to the pressure thickness of the frontal zone

	C number				
	2	3	4	5	6
Mean pressure thickness of zone in mb.	222	224	205	202	180

(VI) Vertical distribution of cloud reported at Bircham Newton in relation to the frontal zone

In order to get a picture of how the reported cloud was distributed in relation to the frontal zone, the percentage frequency of occurrence of cloud at specified levels was calculated for 3 ranges of distance from the front. Table VI shows these results together with the overall percentage frequency. The figures convey the general impression of frontal cloud mainly in the warm air at distances greater than 200 miles but extending downwards through the frontal zone as the surface front is approached. Taking all cases together there is a chance of about 1 in 2 of finding cloud at any point between the lower boundary and a height of 16,000 ft above the upper boundary, the ceiling of course being the 400 mb. level. The two 100% figures in column 3 are due to a solitary case where the upper boundary was 16,000 ft below 400 mb. in that particular range.

Table VI

Percentage frequency of occurrence of cloud at specified levels in relation to the height of the frontal zone (69 cases)

Level	Less than 100 miles from front (23 cases)	100 to 200 miles from front (22 cases)	More than 200 miles from front (24 cases)	All cases combined
16,000 ft	67	67	100	78
14,000 ft	50	67	100	80
12,000 ft	58	60	75	67
Above 10,000 ft	62	55	71	65
Zone 8,000 ft	61	54	44	59
6,000 ft	74	47	47	62
4,000 ft	69	67	42	63
2,000 ft	83	53	50	62
At Upper Limit	87	50	33	58
¾ Depth of Zone	91	64	29	58
½ Depth of Zone	91	46	13	50
¼ Depth of Zone	87	50	13	47
Lower Limit	78	50	21	47
2,000 ft	40	24	17	30
Below 4,000 ft	0	12	0	7
Zone 6,000 ft	0	0	8	3

(VII) Relation of cloud at Bircham Newton to the temperature difference across the frontal zone measured on the saturated adiabatic scale

The difference of saturated potential temperature from the cold to the warm air vertically across the frontal zone is a measure of the thermal strength of the zone. Table VII indicates that there is no correlation between this and the C number.

Table VII

Relation of frontal cloud to the temperature difference across the frontal zone

	C number				
	2	3	4	5	6
Mean Saturated Adiabatic Difference in °F across zone	10.8	10.3	10,7	9.8	11.2

The correlation between Y number and the change of saturated adiabatic potential temperature across the zone was negligible, viz:- .05.

(VIII) Relation of cloud at Bircham Newton to the geostrophic wind shear in the horizontal across the surface front

The range of values in each class was from 10 kt. or less to 60 kt. or more, so that although Table VIII shows an increase in the means from C2 to C5 there can be but small significance in the result. The correlation of wind shear with Y number was likewise small viz:- .17.

Table VIII

Relation of frontal cloud to the geostrophic wind shear across the front

	C number	2	3	4	5	6	All
Wind shear in knots	Mean	28	31	32	34	28	31
	Standard Deviation	16	14	17	22	15	
	No. of cases	13	21	24	13	5	76

(IX) Relation of cloud at Bircham Newton to the rate of increase of geostrophic wind shear in the horizontal across the surface front

The same method of measuring this parameter was adopted as was used by Matthewman - namely the change of wind shear along the normal through Bircham Newton used in the cross-section. The disadvantage pointed out in the original paper, i.e. the difficulty of measuring this shear on the subsequent chart owing to rapid occlusion was avoided by taking only the change during the preceding 6 hours. It was felt too that this might have a greater bearing upon cloud development than the mean rate of change over a 12 hour period. No relationship is revealed. However it is noteworthy that the mean rates are all positive.

Table IX

Relation between frontal cloud and the rate of change of geostrophic wind shear

	C number	2	3	4	5	6	All
Mean rate of increase in knots of wind shear	Mean	1.2	.7	.8	.9	.6	.9
	Standard Deviation	2.2	1.7	1.9	3.5	3.3	
	No. of cases	13	21	24	13	5	76

The correlation of rate of increase of wind shear with Y number was .19; it is doubtful whether this is significant.

/(X)

(X) Relation of cloud at Bircham Newton to the actual wind shear in the vertical across the frontal zone parallel to the surface front

This shear was taken as the difference between the components parallel to the surface front of the winds at the upper and lower boundaries of the frontal zone. Table X shows no correlation whatever between this shear and C number. Likewise the correlation with Y number was zero.

Table X

Relation between frontal cloud and vertical wind shear parallel to front

	C number	2	3	4	5	6	All
Mean shear in the vertical across frontal zone (kt.)	Mean	28	25	20	25	23	24
	Standard Deviation	22	18	17	17	11	
	No. of cases	13	21	24	13	5	76

(XI) Relation of cloud at Bircham Newton to the wind shear in the vertical across the frontal zone normal to the front

The wind shear across the zone normal to the surface front gives somewhat more interesting figures. From Table XI it is seen that there is some increase in the means with increasing C number, but the correlation is only .16. This correlation is improved to .28 if the Y number is substituted.

Table XI

Relation between frontal cloud and vertical wind shear normal to the front

	C number	2	3	4	5	6	All
Mean shear in vertical across zone normal to front (kt.)	Mean	8	11	13	9	19	11
	Standard Deviation	8	9	9	11	13	
	No. of cases	13	21	24	13	5	76

(XII) Relation of cloud at Bircham Newton to the difference in speeds between the surface front and the component of wind normal to the surface front at the upper boundary

Matthewman achieved his highest correlations with this parameter, and Table XII suggests there is a positive relationship also with the C number in the present series of fronts.

Table XII

Relation between frontal cloud and the difference in speeds between warm air and surface front

	C number	2	3	4	5	6	All
Difference in kt. between speed of front and normal component of wind at upper limit.	Mean	3	6	8	7	21	7
	Standard Deviation	15	12	10	12	4	
	No. of cases	13	21	24	13	5	76

/The

The correlation between C number and this parameter is only .23, but if one case is omitted (where the frontal analysis is in some doubt but the wind speeds are very high) this correlation can be improved to .41. Restricting the cases to those of appreciable shear through the frontal zone i.e. where the shear is greater than or equal to 12 kt., the correlation is again slightly improved to .45.

The matter was tested further as it seemed to be the most fruitful sphere of investigation. Still restricting the cases to shear greater than or equal to 12 kt., correlation with the X number was found to be .49. The general correlation with Y number (omitting the one case mentioned earlier) was .37 (68 cases); so that it is clear that the relationship does not arise primarily from an association of both variables with distance from the front.

(XIII) Relation between cloud at Bircham Newton and the difference in speeds between the surface front and the component of wind normal to the surface front at the lower boundary

Nothing very significant is revealed by the following table. The correlation between Y number and this parameter was also very small, namely .14.

Table XIII

Relation between frontal cloud and the difference in speeds between the cold air and surface front

Difference in kt. between speed of front and normal component of wind at lower limit	C number	2	3	4	5	6	All
	Mean	-4	-5	-4	-2	+2	-4
	Standard Deviation	15	8	6	15	10	
	No. of cases	13	21	24	13	5	76

(XIV) Relation between the cloud at Bircham Newton and the ageostrophic motion of the surface front

This parameter is simply the speed of the front (taken as before over a 12 hour period) subtracted from the geostrophic component of wind normal to the surface front measured on the line of cross-section through Bircham Newton. There is some increase in the means from C2 to C6 but obviously little correlation. With Y number the correlation was only .12.

Table XIV

Relation between frontal cloud and the ageostrophic motion of the surface front

Ageostrophic motion of surface front (kt.)	C number	2	3	4	5	6	All
	Mean	13	13	11	16	16	13
	Standard Deviation	8	8	6	14	10	
	No. of cases	13	21	24	13	5	76

It may be of some passing interest to note that the average value of the ratio, speed of front to geostrophic wind normal to front, for the 76 cases is .67.

5. Conclusions

The results are generally in close agreement with those of Matthewman's study (M.R.P.747), and this in spite of the fact that the careful selection of active warm fronts of the present paper represent a very different sample from the heterogeneous collection of January to March, 1942. The fact that the Matthewman cloud classification is closely linked with the distance of the ascent from the surface front in the case of an active warm front has been borne in mind throughout, but the Y classification which does not have this dependence gives essentially the same results as the original classification used by Matthewman.

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The present study confirms the existence of a small but significant correlation (about + 0.4) between the extent of cloud and the component of wind relative to the surface front taken normal to the front at its upper boundary. However it would appear that there are no simple and reliable measurements that a forecaster can make in order to estimate the cloud structure of a warm front.

One result additional to those of the previous investigation is the tendency demonstrated by Table VI for the slope of the edge of the frontal cloud to be steeper than the slope of the front, so that at distances greater than 200 miles from the surface front the cloud lies mainly in the warm air whereas within 100 miles of the front the cloud usually extends down through most of the transitional zone.

References

Matthewman, A.G. 1952. M.R.P. 747, Met. Office, London.

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J = geostrophic wind component perpendicular to AB.

V' = ageostrophic wind component perpendicular to AB.

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At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 800 by Mr. D.H. Johnson.

A copy is available in the Library of the Meteorological Office."

S.C.II/144

24th April, 1953.

AD37929

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The accuracy of 100 mb. contour heights

by
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Summary

Comparison between observed winds and slopes of the 100 mb. pressure surface indicate that the standard error in the measurement of 100 mb. contour height, using radio-sonde data for British land stations, is of the order of 145 feet.

(1) Introduction

100 mb. wind reports show flow patterns at the 100 mb. level to be smoother than those of the troposphere. If, however, we draw the contours of the 100 mb. surface to agree strictly with the measurements of contour height we obtain patterns which are far more irregular and have a much greater variability in space and time than the observed flow. The dominant terms in the expression for the ageostrophic wind depend on the local and spatial accelerations. Since these are small at 100 mb. we can expect the ageostrophic winds to be small. So, knowing the observed winds to be reasonably accurate (1), we must account for the large apparent geostrophic departure in terms of contour height errors. A method of estimating the standard error in measurements of 100 mb. contour height is described below.

(2) The nature of the test

An estimate of the standard error of the contour height measurements can be obtained simply by comparing the measured contour gradients with the observed winds, due consideration being given to errors, other than those of the contour heights themselves, which may affect the comparison. The problem may be formulated as follows. The value which we ascribe to the 100 mb. contour height above a given station is obtained from radio-sonde observations of temperature and pressure by integration of the hydrostatic equation. We know that this value contains not only the usual errors of measurement and computation but also errors due to the fact that the basic observations are not made simultaneously at all levels; they are made over finite intervals of time during which the balloon travels some distance away from the point to which they are taken to refer. We are concerned in the present test only to obtain an estimate of the total error in the values obtained and not to discriminate between the various component parts.

Expression for the standard error

Given two adjacent upper air stations at points A and B.

Then let V = true wind component perpendicular to AB.

U = measured wind component perpendicular to AB.

ε = error in measured component perpendicular to AB.

J = geostrophic wind component perpendicular to AB.

V' = ageostrophic wind component perpendicular to AB.

x = distance measured along AB.

d = distance between A and B.

h = true height of 100 mb. surface.

H = measured height of 100 mb. surface.

Θ = error in measured height of 100 mb. surface.

g, l, have their usual significance. Suffixes A, B denote values of the variables at points A, B,

then $V = V' + J$ where $J = \frac{g}{l}\left(\frac{\partial h}{\partial x}\right)_{100}$

taking mean values over the distance AB for a particular occasion

$$\bar{V} - \bar{V}' = \bar{J} = \frac{g}{ld}\left(h_B - h_A\right)$$

so

$$\left(H_B - H_A\right) + \left(\Theta_B - \Theta_A\right) = \frac{ld}{g}\left(\bar{V} - \bar{V}'\right) \qquad \text{.....(1)}$$

Now let η be the difference between the true wind component at any point along AB and the wind component obtained by linear interpolation between the true winds at A and B.

so that

$$\bar{V} = \tfrac{1}{2}\left(V_A + V_B\right) + \bar{\eta}$$
$$= \tfrac{1}{2}\left(U_A + U_B\right) + \tfrac{1}{2}\left(\varepsilon_A + \varepsilon_B\right) + \bar{\eta}$$

and from (1)

$$\left(H_B - H_A\right) - \frac{ld}{2g}\left(U_A + U_B\right) = \left(\Theta_A - \Theta_B\right) + \frac{ld}{g}\left[\tfrac{1}{2}\left(\varepsilon_A + \varepsilon_B\right) + \bar{\eta} - \bar{V}'\right]$$

Now taking the r.m.s. values of both sides of the equation over a large number, N, of occasions and assuming the Θ, ε, η and V' to be independent, we have

$$\frac{1}{N}\sum\left[\left(H_B - H_A\right) - \frac{ld}{2g}\left(U_A + U_B\right)\right]^2 = \frac{2}{N}\sum\Theta^2 + \frac{l^2d^2}{Ng^2}\left(\tfrac{1}{2}\sum\varepsilon^2 + \sum\bar{\eta}^2 + \sum\bar{V}'^2\right)$$

and the standard error in the measured height

$$= \left[\frac{1}{N}\sum\Theta^2\right]^{\frac{1}{2}} = \tfrac{1}{2}^{\frac{1}{2}}\left\{\frac{1}{N}\sum\left[\left(H_B - H_A\right) - \frac{ld}{2g}\left(U_A + U_B\right)\right]^2 - \frac{l^2d^2}{Ng^2}\left[\tfrac{1}{2}\sum\varepsilon^2 + \sum\bar{\eta}^2 + \sum\bar{V}'^2\right]\right\}^{\frac{1}{2}}$$

Evaluation of the errors

The first term in the expression for $\left(\frac{1}{N}\sum\Theta^2\right)^{\frac{1}{2}}$ has been calculated using 356 pairs of observations for the first nine months of 1952 made by 6 pairs of British radio-sonde stations. The r.m.s. $\left[\left(H_B - H_A\right) - \frac{ld}{g}\left(U_A + U_B\right)\right]$ was 207 feet, implying an upper limit to $\left(\frac{1}{N}\sum\Theta^2\right)^{\frac{1}{2}}$ of 145 feet.

It is reasonable to assume that $\frac{1}{N}\sum\bar{\eta}^2$ is not greater than $\frac{1}{N}\sum\eta_m^2$ where η_m is the error of interpolation at the mid-point of AB. Now an estimate, 49 kt.^{2}, of $\frac{3}{2N}\sum\varepsilon^2 + \frac{1}{N}\sum\eta_m^2$ has previously been obtained (1) by comparing an interpolation between winds at Aldergrove and Downham Market

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with the wind observed at Liverpool. This estimate applied to a winter sample of winds and the distance Aldergrove - Downham exceeds the distance between any of the pairs of stations used in the present test. So we may safely take the value of the terms $\frac{1}{2N}\Sigma\varepsilon^2+\frac{1}{N}\Sigma\bar{\eta}^2$ as being less than 49 kt^2.

$\frac{1}{N}\Sigma\overline{V'}^2$ will be of the same order as $\frac{1}{N}\Sigma(V_A')^2$ or $\frac{1}{N}\Sigma(V_B')^2$. Now representative values of the observed time and space variations of 100 mb. wind are 11 kt./12 hr. and 12 kt./300 mi. respectively in winter and less in summer. We can therefore deduce from the relation between acceleration and geostrophic departure that the ageostrophic wind is likely to average 3 or 4 kt. 25 kt^2. is probably a generous allowance for the mean square of the ageostrophic wind components at A or B and so for $\frac{1}{N}\Sigma\overline{V'}^2$.

The factor ld/g combined with that required to convert kt. to ft./sec. varies between different pairs of stations from 4 to 5 sec. From these considerations we can estimate that the term $\frac{\ell^2d^2}{Ng^2}\left[\frac{1}{2}\Sigma\varepsilon^2+\Sigma\bar{\eta}^2+\Sigma\overline{V'}^2\right]$ will be less than 1850 ft^2. So $\left(\frac{1}{N}\Sigma\theta^2\right)^{\frac{1}{2}}$ will be greater than

$\frac{1}{2^{\frac{1}{2}}}\left[(207)^2-1850\right]^{\frac{1}{2}}$ = 142 ft.

Strictly, the values of U employed in these calculations should refer to points vertically above A and B at 100 mb., but the spatial variation of U can reasonably be expected to be so small on the average over the 10 to 30 miles the balloon is usually carried away from its station as to have an insignificant effect on these results.

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It is of interest to compare the 100 mb. standard error with those given by Murray (2) for lower levels, which are contained in Table I.

Table I. Standard errors in measurements of heights of isobaric surfaces at British land stations.

Pressure level, mb.	700	500	300	200	100
Approximate height, ft.	10,000	18,000	30,000	38,000	53,000
Standard error, ft.	20	30	70	100	145

Above 500 mb. the standard error increases linearly with height.

References.

(1) Johnson, D.H. 1953. Met. Mag., Vol. 82, p. 44.

(2) Murray, R. 1951. M.O.21 Technical Note No. 37, Met. Office, Dunstable.

$$= \tfrac{1}{2}(U_A + U_B) + \tfrac{1}{2}(\varepsilon_A + \varepsilon_B) + \eta$$

and from (1)

$$(H_B - H_A) - \frac{\ell d}{2g}(U_A + U_B) = (\Theta_A - \Theta_B) + \frac{\ell d}{g}\left[\tfrac{1}{2}(\varepsilon_A + \varepsilon_B) + \bar{\eta} - \bar{V}'\right]$$

Now taking the r.m.s. values of both sides of the equation over a large number, N, of occasions and assuming the Θ , ε , η and V' to be independent, we have

$$\frac{1}{N}\Sigma\left[(H_B - H_A) - \frac{\ell d}{2g}(U_A + U_B)\right]^2 = \frac{2}{N}\Sigma\Theta^2 + \frac{\ell^2 d^2}{Ng^2}\left(\tfrac{1}{2}\Sigma\varepsilon^2 + \Sigma\bar{\eta}^2 + \Sigma\bar{V}'^2\right)$$

and the standard error in the measured height

$$= \left[\frac{1}{N}\Sigma\Theta^2\right]^{\frac{1}{2}} = \tfrac{1}{2}^{\frac{1}{2}}\left\{\frac{1}{N}\Sigma\left[(H_B - H_A) - \frac{\ell d}{2g}(U_A + U_B)\right]^2 - \frac{\ell^2 d^2}{Ng^2}\left[\tfrac{1}{2}\Sigma\varepsilon^2 + \Sigma\bar{\eta}^2 + \Sigma\bar{V}'^2\right]\right\}^{\frac{1}{2}}$$

Evaluation of the errors

The first term in the expression for $\left(\frac{1}{N}\Sigma\Theta^2\right)^{\frac{1}{2}}$ has been calculated using 356 pairs of observations for the first nine months of 1952 made by 6 pairs of British radio-sonde stations. The r.m.s. $\left[(H_B - H_A) - \frac{\ell d}{g}(U_A + U_B)\right]$ was 207 feet, implying an upper limit to $\left(\frac{1}{N}\Sigma\Theta^2\right)^{\frac{1}{2}}$ of 145 feet.

It is reasonable to assume that $\frac{1}{N}\Sigma\bar{\eta}^2$ is not greater than $\frac{1}{N}\Sigma\eta_m^2$ where η_m is the error of interpolation at the mid-point of AB. Now an estimate, 49 kt.^{2}, of $\frac{3}{2N}\Sigma\varepsilon^2 + \frac{1}{N}\Sigma\eta_m^2$ has previously been obtained (1) by comparing an interpolation between winds at Aldergrove and Downham Market